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PLATE TECTONICS

Developments in Geotectonics 7

EAST AFRICAN RIFTS

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FOREWORD

Papers in this special issue of *Tectonophysics* were presented at a one day Upper Mantle Committee Symposium on "*Rifting in Africa*", held on August 7, 1971 as part of the 15th General Assembly of the International Union of Geodesy and Geophysics in Moscow. The International Upper Mantle Committee (I.U.M.C.) was a Committee set-up jointly by the International Union of Geodesy and Geophysics and the International Union of Geological Sciences to coordinate the Upper Mantle Project, an international programme of research on the solid earth sponsored by the International Council of Scientific Unions.

Throughout the Upper Mantle Programme the East African rift system has been a subject of attention for scientists from many parts of the world and at the end of 1970 the chairman of I.U.M.C., Prof. V.V. Beloussov, asked me to organize a one day symposium to be held in Moscow as part of the 15th General Assembly of I.U.G.G. Professor E.E. Milanovsky was nominated to help on the part of the Soviet Union and I would like to record my appreciation for all his efforts with the local organisation. Twenty scientists were contacted with a view to presenting papers at the symposium and it is a measure of the enthusiasm for the East African rift system that we ended up on the day of the meeting with twenty-two papers! The support for the meeting was very gratifying to the organisers who extend their thanks to all the participants for their part in making the meeting a success.

Following the pioneering works of J.W. Gregory, it has become customary to think of the East African rift system as including the Red Sea and Gulf of Aden as well as the rifts in East Africa. A three day meeting on the Red Sea and Gulf of Aden had been held in London in March 1969 under the auspices of the Royal Society (*Philos. Trans R. Soc. Lond.* A267 (1970)). In view of this and the fact that the U.M.C. Symposium in Moscow was limited to one day, it was decided to restrict this meeting to a discussion of the rifts in East Africa. Since the London Symposium, much work has been carried out on the "Afar triangle" of Ethiopia and it was obviously desirable to have further papers and discussions on this region. The symposium therefore started where the London symposium left-off, by including the rifting in Africa southwards from Ethiopia.

It will be noticed from the first seven papers that the Afar region remains controversial. Various interpretations of the geological structure, volcanicity, gravity and seismic results are still out of harmony and in need of further appraisal. As more data are acquired, it is becoming more apparent that much of the crust here is unusual and somewhat similar to that of Iceland. Among the succeeding papers, readers will find a considerable amount of support for the idea that the rift in East Africa represents the early stages in the break-up of a continental plate or to use McCall's delightful phraseology represents an "aborted ocean". However, it should be remembered that compared with the Atlantic and Indian Oceans, its history is of the order of 20 m.y. rather than 200 m.y. Also, there seems to be

growing evidence for the go–stop–go nature of its evolution. Not all the contributors agree and readers will find some interesting dissenting views.

It was hoped that the manuscripts of the papers presented at the meeting would all be submitted by November 30, 1971. In fact, they were received in Newcastle between October 1971 and January 1972 and were all handed over to the publisher on February 1, 1972. The papers form U.M.P. Scientific Report no.40 which is followed by the Proceedings of the final review Symposium (U.M.P. Sci. Rep. 41, *Tectonophysics*, 13) of the U.M.P. held in Moscow 9–13th August 1971, the Project having terminated at the end of 1970.

The symposium received financial support from I.U.G.G., I.U.G.S. and UNESCO which helped to contribute to the travel expenses of several of the participants and for which we are sincerely grateful.

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SURFACE STRUCTURE AND PLATE TECTONICS OF AFAR

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ABSTRACT

Mohr, P.A., 1972. Surface structure and plate tectonics of Afar. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15 (1/2): 3–18.

The surface structural elements of Afar are indicative of crustal tension. The dominating feature is intense slicing of the Afar Neogene volcanics by early Pleistocene fault-belts, whose trends are related to those of the three rift systems converging on Afar. This faulting shows that crustal extension affected the whole width of Afar, rather than being confined to oceanic-type, axial zones. Nevertheless, the computed amounts of extension agree well with results of plate analysis of the Afar triple-junction. Refined plate analysis for this region yields data consistent with rotation of the Danakil horst, with inferred presence of cross-rift faults, with longitudinal shearing in eastern Afar, with the apparent asymmetry of spreading in the western Gulf of Aden, and with an early episode of north–south spreading in southernmost Afar. Crustal spreading in Afar seems to be a long-term stop-go phenomenon, initiated at the beginning of the Cainozoic (70 m.y. B.P.). The largest tectonic episode was probably related to the Upper Pliocene–Lower Pleistocene swell uplift of Ethiopia, and the splitting apart of the Danakil and Aisha horsts.

INTRODUCTION

The eastern rift system of Africa terminates in the north where it meets the Red Sea and Gulf of Aden at the Afar triple junction (Baker et al., 1972). This junction area lies largely above sea-level, and its additionally distinctive structural and volcanic geology all suggest caution in interpreting Afar as typical of incipient oceanic development. On the contrary, it may be as typical of such development as Iceland is typical of the Mid-Atlantic ridge.

GEOLOGICAL SETTING

Sea-floor spreading is indicated to be operative in the Red Sea (half spreading rates of 1.0–1.6 cm/year: Vine, 1966; Phillips and Ross, 1970) and Gulf of Aden (0.9–1.1 cm/year: Laughton et al., 1970). McKenzie et al. (1970) have used these and other data to make a geometric analysis of the Afar triple junction, and their results confirm the thesis of Laughton (1966) that Afar must be a region of new crust (but not necessarily oceanic-type crust – see Dakin et al., 1971) generated in response to drift of Arabia from Africa

since the Miocene (20 m.y. B.P.).

Geological observations indicate a certain complexity in the junction area (Mohr, 1967; Bannert et al., 1970; Tazieff, 1970). Afar is almost completely circumscribed by continental crustal blocks, and lacks a narrow, oceanic-type spreading zone except in the extreme north. The sialic Danakil and Aisha horsts separate Afar from the Red Sea and Gulf of Aden, though Hutchinson and Engels (1970) consider the Danakil horst to be the eastern, exposed rim of a tilted sialic block which extends beneath the whole of northern Afar. The faulted, warped and injected margins of the Ethiopian and Somalian plateaux, overlooking Afar from west and south respectively, expose sialic rocks for as much as 50 km inside the main escarpments, confirming the revisions of Freund (1970) and Mohr (1970a) to the plate analysis of McKenzie et al. (1970).

STRUCTURE

Afar floor

In northern Afar a series of Neogene and Quaternary sandstones, limestones and thick evaporites (with minor marginal volcanics) is partially overlain by products of an axial zone of active basaltic shield volcanism (Brinckmann and Kürsten, 1969a; Barberi and Varet, 1970). In central and southern Afar, extensive (?) Lower Miocene–Upper Pliocene terrestrial flood basalts (Afar Series) are overlain by silicic piles, frequently with calderas, by restricted Quaternary flood basalts and rather extensive Quaternary lacustrine and fluvial sediments (Taieb, 1971). Peralkaline granites of Oligocene–Quaternary age pierce this succession (Brinckmann and Kürsten, 1969a, b), notably in west-central Afar and on the Danakil horst.

The floor of Afar shows an intense structural deformation, dominated by a Lower–Middle Pleistocene episode of normal faulting which affected virtually the whole width of the depression (Mohr, 1967, 1970b). Belts of faulting extend from the northern apex of Afar, by both right en echelon displacements and curvilinear adjustments, through the Erta-ali and Alayta volcanic shields before widening out and bifurcating into central Afar (Fig.1 and 2). The individual faultbelts are typically 10–30 km across, and persist for more than 100 km length before bifurcating (e.g., around older nuclei), merging, abruptly changing trend, or dying out. The comprising faults tend to be regularly spaced, commonly 200–500 m apart, and average throws are 20–100 m. Throw directions are usually sympathetic (ratchet faulting), but can be mixed (horst–graben faulting). The disposal of ratchet faulting in Afar suggests relative uplift of the older crustal nuclei.

The Afar fault-belts tend to be orientated parallel to the structural direction of the nearest converging rift system. Complications occur near Lake Abbe (the triple-point), and also in east-central Afar where additional crustal stresses have been induced by rotation of the Danakil horst out of Afar (Laughton, 1966; Burek, 1970; Mohr, 1970b). Where intersecting or partially non-adapting fault-belts occur, it appears that the faults

concerned were closely synchronous (Mohr, 1967; Tazieff, 1971), indicating the action of a single stress-field.

Although Red Sea tectonics affect Afar via the Salt Plain, and Gulf of Aden tectonics via the Gulf of Tajura, crustal extension is notably distributed over the whole width of Afar rather than restricted to hypothetical neo-oceanic axial zones. Only in the case of southwestern Afar is there a tendency for an axial volcano–tectonic lineament (Wonji fault belt) to extend out from the rift, and this narrowing of the zone of tectonism has left extensive regions of southern Afar atectonic and blanketed by thick lacustrine sediments (Mohr, 1967; Taieb, 1972). The projection of the Wonji fault belt well into Afar enables a triple point to be assigned in the Lake Abbe region.

Afar margins

The early history of Afar is preserved only in the continental crustal margins, a fact which of course is entirely compatible with the sea-floor spreading hypothesis. Mesozoic marine sediments, lying upon Precambrian schists (Mohr, 1962a), thicken from the plateau into the downwarped Afar marginal zone (Jepsen and Athearn, 1962; Mohr and Gouin, 1967). The Palaeogene trap series basalts are likewise warped and thicken riftwards across the margin; there is no evidence for a directional migration of the feeder dyke-swarms with time, as has been postulated for the Icelandic rift by Gibson (1967). It should be noted that Tazieff (1970), Barberi and Varet (1970) and others deny the existence of any warping of the Ethiopian plateau–Afar margin – but see the detailed mapping of Mohr (1971b).

The warping and faulting of the Afar-plateaux margins have been intimately related to the pulses of Ethiopian swell uplift (Mohr, 1967; Baker et al., 1972). The latest, Upper Pliocene–Pleistocene uplift was the largest in terms of vertical magnitude, and may prove to have been closely synchronous with antithetic and graben faulting of the margins, as well as with the floor fault-belts of Afar. Despite building up and/or uplift of the Afar crust above sea-level, the margin zones and plateaux have been uplifted faster, as indicated by raised lake-levels (M. Taieb, personal communication, 1970; Mohr, 1971b) and by upwarping of the plateau rims (Mohr, 1962b; Mohr and Gouin, 1967; Tazieff, 1970).

The Ethiopian Plateau–Afar margin is 30–50 km wide, confined between the outer structural margin (the faulted plateau rim) and inner structural margin (faulting of sialic rocks against Afar floor strata). The margin is transected in places by severe cross-faulting, folding and tight monoclinal warps (Mohr, 1967; Arkin et al., 1971), a feature also of the narrower Somalian Plateau–Afar margin. Dyke swarms tend to parallel and be concentrated along zones of more severe margin warping, which are not everywhere coincident with subsequent zones of marginal faulting (Mohr, 1971b).

CRUSTAL EXTENSION IN AFAR

The Afar floor fault-belts are considered to be essentially Lower–Middle Pleistocene in

age, despite Holocene tectonism in the Erta-ali and Wonji fault zones (Mohr, 1970b; Taieb, 1971, 1972). There is not yet recognition of any Tertiary crustal extension episodes in Afar, though the Afar Series flood basalts must be related to some such episode.

On the basis of average data for the Afar fault-belts, the amount of Pleistocene crustal extension can be estimated. Assuming 2–4 faults/km in a 25 km-wide belt, an average throw of 75 m and a hade of 30°, and using the data of Fig.1 and 2 the values in Table I were obtained.

TABLE I

Distension data for Afar fault belts

Region		Distension direction	Distension amount (km)
North-central Afar	(12½°N)	NE–ENE	25
Eastern Afar	(42°E)	NNE	20
Southern Afar	(10½°N)	ESE	5

These three values relate in remarkable consistency with the spreading vectors obtained by plate analysis of the Afar triple junction (McKenzie et al., 1970). It confirms the influence of all three rifts on the tectonics of Afar, not solely the Red Sea (Tazieff, 1970), and indicates a Quaternary triple-point near Lake Abbe (11°N 41½°E). The apparent smooth transition between Red Sea and Gulf of Aden tectonics in eastern Afar in fact relates to Danakil horst rotation, and a tempting correlation with Carlsberg Ridge and Gulf of Aden tectonism on the grounds of parallelism alone (Tazieff, 1971) should be avoided, seeing that quite different crustal plate-pairs are involved.

The episodic nature of Afar crustal extension, allied to its distribution over the whole floor of Afar, makes an interesting contrast with the ocean ridges. It seems that rupture of the Afar crust occurs rather rarely (stop–go), but when rupture does occur concomitant with swell-uplift it is massive over an extensive area. This indicates the presence of a thin, brittle upper crust underlain by a thick, stiffly viscous lower crust undergoing uniform extension. There has been no narrowing of the zone of tectonism with time, as has occurred in parts of the Eastern Rift of Africa (Gibson, 1969; Baker et al., 1972).

A transform fault has been proposed by Gass et al. (1972) in the Red Sea east of the Danakil horst. Whether embryo or ill-developed transform faults occur in the broader spreading-zone of Afar remains uncertain, but the presence of cross-rift lineaments (Mohr, 1967) suggests a type of quasi-transform behaviour.

AFAR AND PLATE TECTONICS

A qualitative geometric analysis of Afar in terms of plate evolution has been attempted

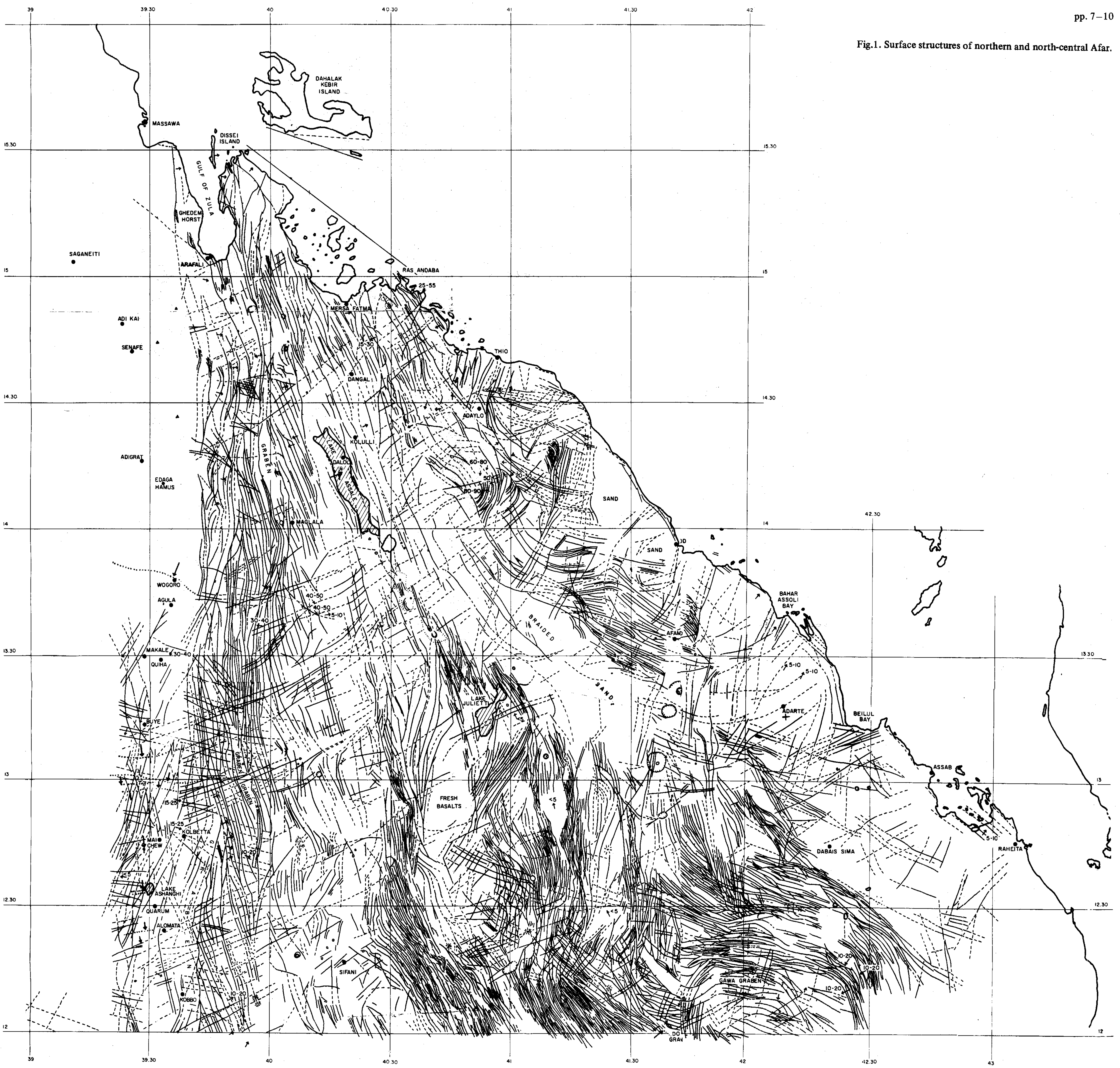

Fig.1. Surface structures of northern and north-central Afar.

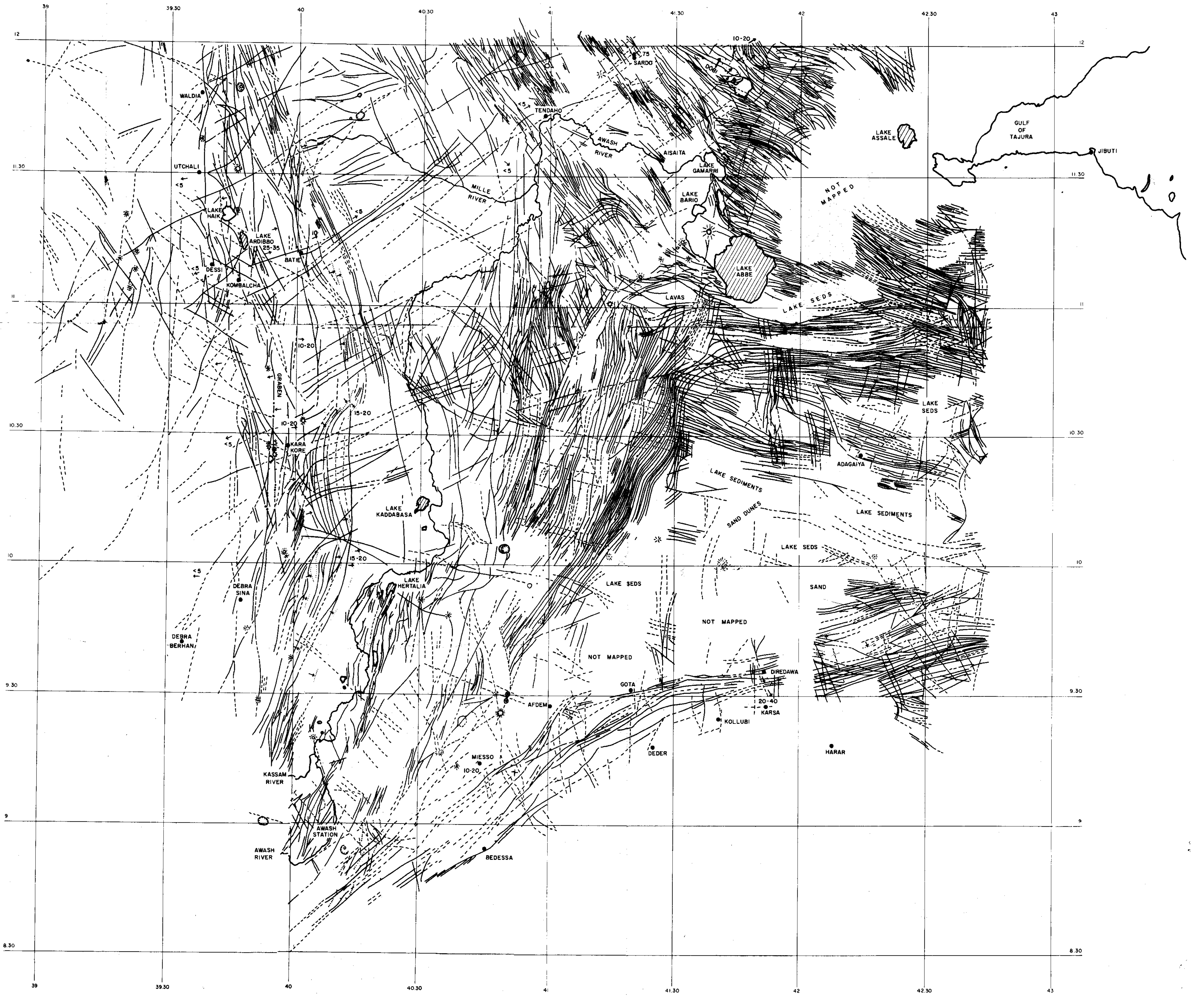

Fig.2. Surface structures of southern and south-central Afar.

by Mohr (1970b). It was assumed that the greater part of the Red Sea and Gulf of Aden floors is neo-oceanic crust, such that Afar too is essentially lacking in continental crust (McKenzie et al., 1970; but see also the pertinent reservations and conclusions of Dakin et al., 1971). The Danakil and Aisha horsts were included as separate microplates in the analysis. It remains to be proved whether plate analysis on such a small scale is justified, but if it fails it is important to see where and why in terms of the consequences for structure of continental rifts.

Fig.3 shows a proposed model development of the opening of the Red Sea and Gulf of

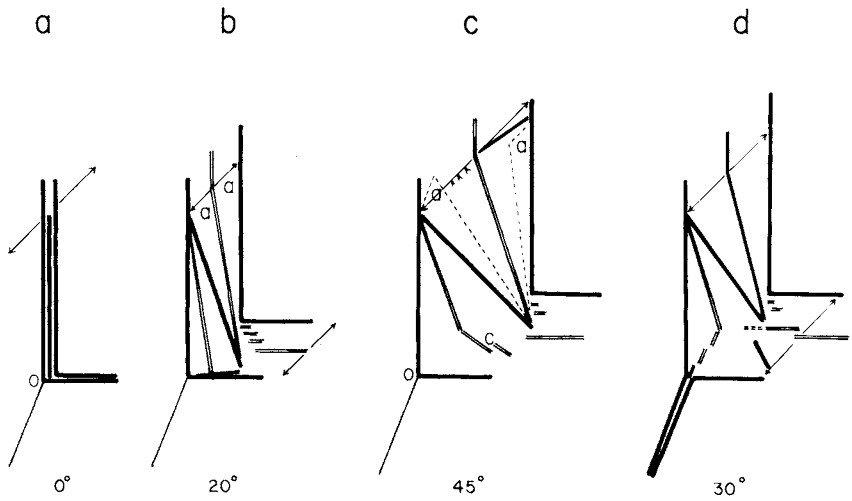

Fig.3. Model for plate evolution of Afar triple-junction.
a: Initial scheme (early Cainozoic).
b: 20° rotation of Danakil horst. Note: (1) sinistral sliding of horst termini (north end assumed fixed here for simplicity's sake); (2) necessary northerly displacement of Gulf of Aden spreading axis to be able to generate new crust in the western reaches; (3) an east–west gap between field of horst-rotation and Somali block to south; (4) maintaining the southern termination of Red Sea spreading at the corner of the Arabian block leads to more severe internal plate deformation than in the presented model.
c: 45° rotation of Danakil horst (horst remaining undivided). Note: (1) overlapping of sector *a* generated at 20° rotation, across Red Sea spreading vector northeast of horst. Conversely, in the eastern Red Sea a transverse paar would be formed (but further rotation would begin to close this again). See text for discussion of how overlapping (and distension) are accomodated longitudinally throughout the plates concerned; (2) if Afar spreading terminates southward at a line of Red Sea spreading vector extending from point *O* (the Lake Abbe triple-point lies on such a line projected from the mouth of the Ethiopian Rift), then plate overlapping in Afar would be accomodated by longitudinal compression (see text) and/or by re-orientation of spreading axes in eastern Afar to position *c*.
d: 35° rotation of Danakil horst, with splitting-off of Aisha horst at 30° position. Note: (1) triple-point develops with the superpositioning of young African rift tectonism on older structures of southern Afar; (2) this tectonism involves appreciably more rapid crustal extension than for the preceding evolution of the African rift; (3) the Gulf of Aden spreading axis is displaced north beginning at the intercept of the spreading vector from the southeastern end of the Aisha horst. This is observed (Laughton et al., 1970).

Aden with an intervening, rotating sliver of continental crust (Danakil–Aisha horst). Sea-floor spreading occurs on both sides of the rotating block, with spreading poles situated near the south end of the block for the eastern, Red Sea zone, and the north end for the western, Afar zone. In this manner the total spread of the two zones is kept equal to that of the single Red Sea axis farther north. Of necessity, the two spreading axes (or zones) rotate at half the angular velocity of the rotating block.

Development of this model leads to some interesting corollaries:

(1) There is secondary but significant longitudinal plate collision at the "open" ends of the bifurcated spreading zones (Note: other phenomena such as migration of the rotation-poles and deformation of the spreading lines are additionally possible.) Such collision would be expected in the western Red Sea (about latitude 16°N) and in east-central Afar (Fig.3). But rather than consumption of crust at a precisely located trench paralleling the Red Sea spreading vector, it seems more probable that minor plate collision involving new, thin lithosphere will lead, in the model proposed, to longitudinal shuffling of the crustal slices formed by extension and normal faulting..

In the southern Red Sea, a single spreading axis is indicated by both bathymetric and volcanological evidence (see Gass et al., 1972). In Afar, the wide zone of extension has produced fault-belts which might easily accommodate small increments of longitudinal compression by flexing and allied transcurrent shuffling of individual faulted crustal slices. Fig.3 indicates that such transcurrent motions would be dextral in the southwestern Red Sea, and sinistral in northeastern Afar. The latter is confirmed by the observations of Dakin et al. (1971) and Mohr (1971a). It must be emphasised that such transcurrent movement is greatly subordinate in magnitude to the normal faulting of crustal extension. Other possible secondary plate phenomena are shown in Fig.3c.

(2) The termini of the rotating block slide relative to the main continental blocks (Africa and Arabia), in a manner dependent on the vector of Red Sea spreading and the orientation of the continental margins. In the model presented, one or both termini of the Danakil horst will slide sinistrally past the African and Arabian blocks, until perpendicularity with the Red Sea spreading vector is reached, when the sense flips. Terminus shear should be detectable in seismic focal mechanisms for the Gulf of Zula–Salt Plain and Bab el Mandeb regions.

(3) The sliding of the southern terminus of the Danakil horst is such that in the northwestern corner of the Gulf of Aden, spreading there must be accommodated by a northwards translation of the Gulf spreading axis (Fig.3). In fact, the spreading axis of the western Gulf of Aden does occur well north of the median line (Laughton et al., 1970).

(4) The role of the Aisha horst in the evolution of Afar remains uncertain, lacking geological and geophysical constraints. But if plate theory is valid on this scale, then rather than being a protruberance on the Somalian plateau, it seems necessary to migrate the horst out from an origin in the Ethiopian rift funnel. Consequently, a minor east–west spreading axis/zone in southernmost proto-Afar is required.

In Fig.3d a model is proposed for the detachment of the southern end of the Danakil horst to form the Aisha horst, by means of westerly extension of Gulf of Aden spreading.

After detachment along the Gulf of Tajura suture, the Aisha horst appears to have taken no further part in the rotation of the Danakil horst, and this enables a date for its detachment to be calculated. Since the offset of parallel shorelines of the two horsts is 70–80 km, and assuming a 3 cm/year eastwards motion of the southern end of the Danakil horst relative to Africa, an initial splitting apart is required at 2½ m.y. B.P. This in turn gives a lower age-limit to the Gulf of Tajura and to the Afar fault-belts, as well as to the northwards extension of African rift tectonism across southern Afar.

(5) Constraints on the plate model come from consideration of the angular opening rate of northern Afar. Red Sea-trend faults in Afar terminate some 450 km southwest of the presumed rotation pole situated near the Gulf of Zula. If the integrated Red Sea spreading rate of 3 cm/year applies to central Afar, according with the model in Fig.3, then the average angular opening rate has been 0.014 sec of arc/year. An independent estimate can be derived from the generally accepted view that the Africa–Arabia split was initiated about 20 m.y. B.P. As the median Danakil horst block lies at approximately 30° to the outer margins of the Red Sea trough (see also Burek, 1970), an average angular rotation rate of 0.005 sec of arc/year is obtained.

These two angular opening rates are discrepant by a factor of three. Either the spreading rate in Afar is slower than in the Red Sea farther north (in which case plate theory breaks down for the entire Red Sea–Gulf of Aden region), or the Afar pole of rotation lies well within the Salt Plain (structural evidence is unfavourable to this concept), or spreading was initiated at the beginning of the Cainozoic.

The geometric plate analysis of Afar offered here is only one of several possible models (Mohr, 1971c). More geological and geophysical data are required before further refinement can be attempted. If plate tectonics proves to be valid on this scale, the next task is to correlate the known evolution of the Gulf of Aden (Laughton et al., 1970) with that of Afar. Of particular interest would be a search in Afar for events corresponding to the change in Gulf of Aden spreading behaviour 10 m.y. B.P. Meanwhile it is evident that there was an important change in Afar evolution in the early Quaternary, with a subsequently greater role being played by the African rift system at this triple-junction.

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VOLCANISM IN THE AFAR DEPRESSION: ITS TECTONIC AND MAGMATIC SIGNIFICANCE

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ABSTRACT

Barberi, F., Tazieff, H. and Varet, J., 1972. Volcanism in the Afar depression: its tectonic and magmatic significance. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 19–29.

Three main types of volcanism, viz. oceanic, continental and stratoid, reflect the tectonic evolution of Afar.

Stratoid volcanism (flood basalts and ignimbrites) is believed to be an early manifestation of continental rifting, as exemplified by the Ethiopian Rift. It probably developed during the early stages of Red Sea and Gulf of Aden rifting.

Continental volcanism is located on either side of the rift's axis; it is mainly silicic with minor proportions of basalts. The large, essentially silicic, central volcanoes present along the margins of the depression north from 11°00′N are believed to be due to interactions between subcrustal magma and sialic crust.

Volcanism, of oceanic type, essentially basaltic with differentiates reaching to peralkaline silicics, is related to crustal separation. It occurs only along the rift-in-rift axes of the Red Sea–Gulf of Aden megastructures and is partly superimposed on the stratoid volcanics. The southern edge of this oceanic megastructure marks the northern limit of the continental rift (main Ethiopian rift). Tectonic and thermal evidence suggests the rise of upper mantle material beneath the active axes of the Afar depression.

TECTONICS

Sufficient field data are now available for the following broad statements about the general structure of Afar and the nature of the so called triple junction of the Red Sea, Gulf of Aden, and Ethiopian Rift to be proposed:

(1) In the area investigated (Fig.1), the Afar depression exhibits only tensional faults and fissures, usually disposed en-echelon. Some structures, such as the curvilinear grabens of central-eastern Afar have been considered in the past as due to rotation (Mohr, 1968) or shear (Mohr, 1970). Field evidence, however, does not support this interpretation for the following reasons: (1) rhyolitic centres affected by these faults do not show any detectable horizontal displacements; and (2) all observed faults are tensional.

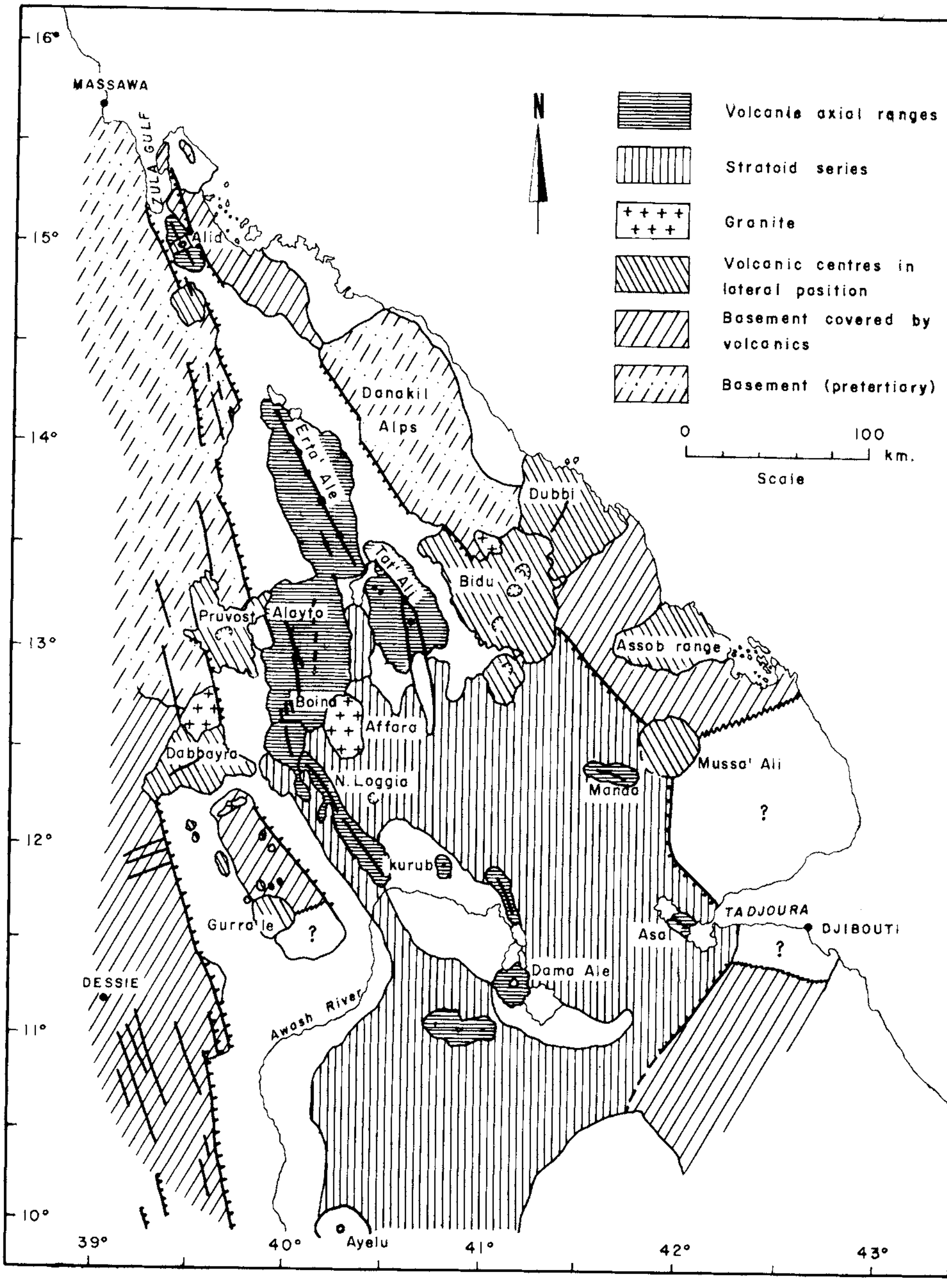

Fig.1. Simplified geological map of the Afar depression showing the different volcanic units.

(2) The northern apex of the Afar depression (from latitude 15°N to 12°N) belongs to the Red Sea NNW–SSE trend. It shows the Red Sea tectonic rift-in-rift axial structure, offset en-echelon to the west where the main central trough of the Red Sea dies out off Massawa (Barberi et al., 1970; Gibson and Tazieff, 1970) (Fig.2).

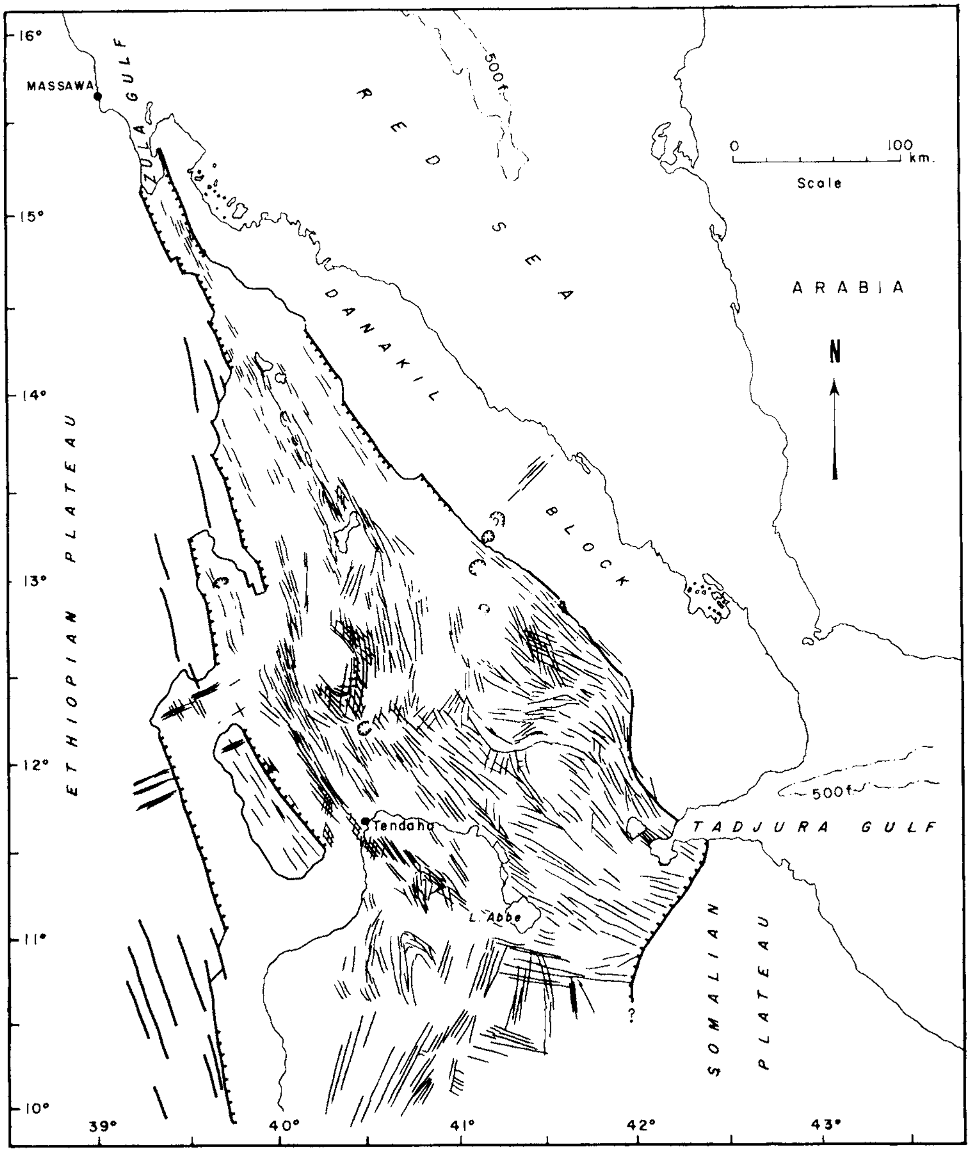

Fig.2. Simplified tectonic map of the Afar depression.

(3) The outline of the faulted western scarp of the depression reflects the en-echelon disposition of the axial grabens of northern Afar, whilst that of the eastern margin (Danakil horst) is defined by rectilinear faults of Red Sea trend. This geometry suggests the breaking up of continental crust. It is consistent with the surface of the depression being formed partly by lavas of oceanic character erupted along the rift axes between sialic crustal blocks, and partly by lavas formed by melting or assimilation of sialic crust. Structural deviations from the main NNW–SSE trend are thought to have been controlled either by pre-existing structures within the underlying basement, or in the case of curvilinear faults, by underlying Miocene granites. Such an interference fault-pattern is obvious around the Affara granite batholith (Fig.1 and 2).

(4) Apparently, the East African rift system appears in the extreme southwest corner of the Afar triangle only; this represents less than one quarter of its total area (Fig.2). It is marked by SSW–NNE faults which disappear in the Lake Abbe-Tendaho area almost as soon as they meet the Red Sea and Gulf of Aden trends (Tazieff, 1971). Afar proper, as tectonically defined, can therefore be *clearly distinguished* from the Ethiopian Rift.

(5) Therefore, the depressed area called "the Afar triangle" should be divided into two quite different structural units: the Red Sea – Gulf of Aden megastructure (NNW–SSE and WNW–ESE trend respectively), of typically mid-oceanic ridge type, and the northern end of the continental Great Rift Valley of East Africa (NNE–SSW).

(6) Although the Gulf of Aden tectonic trend has in the past been considered to be ENE–WSW, i.e., parallel to the Arabian and Somalian coast lines, recent morphological and magnetic maps (Laughton et al., 1970) of the axial Sheba ridge indicate the main trend, clearly visible east of longitude 57°E, to be WNW–ESE. This orientation also characterizes the main tectonics on both sides of the Gulf of Aden as mapped by Azzaroli and Fois (1964) and by Beydoun (1970). At the western end of the east–west oriented Gulf of Tadjoura trench, the Shebaean trend reappears in the Ghoubbet-al-Kharab.

(7) There is no field evidence for an ENE–WSW spreading axis extending inland from the Gulf of Tadjoura through the Lake Abbe (Fairhead and Girdler, 1970; McKenzie et al., 1970; Mohr, 1971). On the contrary, the whole area is split by NW–SE open fissures with horst and graben structures. They all abruptly disappear against the NE–SW scarp which defines the edge of the Somali plate in the Ali Sabieh region.

(8) Over several thousands of square kilometres, central Afar is characterized either by imperceptible merging of these Red Sea and Gulf of Aden trends (on average in a NW–SE direction), or by a criss-cross pattern of conjugated sets of faults (NNW–SSE and NW–SE).

(9) These observed facts lead us to consider the Gulf of Aden and Red Sea rift systems as belonging to a single megastructure (Tazieff et al., 1972), as opposed to the concept of the Afar representing a "funneling out" of the African Rifts system (Mohr, 1967; Baker, 1970).

(10) The aeromagnetic survey of Afar (Girdler, 1970; Girdler and Hall, 1972, this volume) as well as the gravimetric anomalies (Makris et al., 1972, this volume) support this hypothesis. Strong positive gravity anomalies are found beneath the axial ranges in the

investigated area, and magnetic anomalies show a progressive transition from an E–W direction in southern Afar to a NNW–SSE direction in northern Afar. There is no evidence that the Wonji fault belt carries on North of Lake Abbe to the Red Sea. The ENE alignment of the Bidu calderas most probably represents a tensional volcano-tectonic structure affecting the Danakil horst only and not the depression. There are grounds for believing this direction to be a pre-rift structural feature in the basement (Tazieff et al., 1972).

(11) The tectonically and volcanically active, axial grabens in the Afar depression, are characterized by marked, sometimes exceptionally swift, vertical uplifts. The axial trough of northern Afar is dotted by a series of volcano-tectonic horsts (Barberi and Varet, 1970); in central Afar, the Asal graben exhibits beautiful uparching; oyster beds, presently located 100 m above sea level in the relatively down-faulted graben axis, have yielded a radiocarbon age of 5,880 ± 150 years B.P., i.e., an average rate of uplift about 2 cm per year. The spectacular fissuring produced by this uparching accounts for active seismicity and volcanism linked with tectonic block separation. Opening of new gaping fissures, accompanied by strong earthquakes in the Asal area, is evidence for spreading. In these axial grabens, the uplift of the rift shoulders is not concomitant with a simple subsidence of rift floor, as classically proposed for "normal" rift valleys. The obviously uprising central floor is merely upheaved at a slower rate than the rift shoulders. This uplift is probably related to the bulging of upper mantle material beneath the graben, by a process which could be similar to that outlined by Gass (1970) for the Red Sea.

(12) The proposed interpretation of the Afar tectonics implies anticlockwise rotation of the Danakil horst; its southern part perhaps is still somewhat linked to the Arabian plate or, in any case, the severing is comparatively small; the separation from the Nubian plate is equally small at the horst's northern extremity (Zula area). The angle of rotation probably does not exceed $18^\circ \pm 10^\circ$ (Tazieff et al., 1972), which is smaller than Burek's estimates (1970); the exact angle is hardly determinable on a purely geological basis. Geophysical work under way should lead to a more precise picture.

VOLCANISM

A fairly coherent picture of the Afar volcanism is emerging from field relationships and some 40 radiometric age determinations obtained from granitic, volcanic and sedimentary rocks (Barberi et al., 1972). Remarkable age concordances indicate the Lower Miocene (i.e., 25 m.y.) as the probable period of initiation of the Afar Rift.

Subaqueous volcanism is important in Afar, mainly along a band extending from the Gulf of Zula to the Gulf of Tadjoura (Fig.3). Migration of volcanically active zones as well as evolution of the eruptive types since the beginning of the rift formation have produced a mosaic of volcanic products of various ages and nature, locally interbedded with clastic sediments (as near the Awash basin) or evaporites (as in northern Afar).

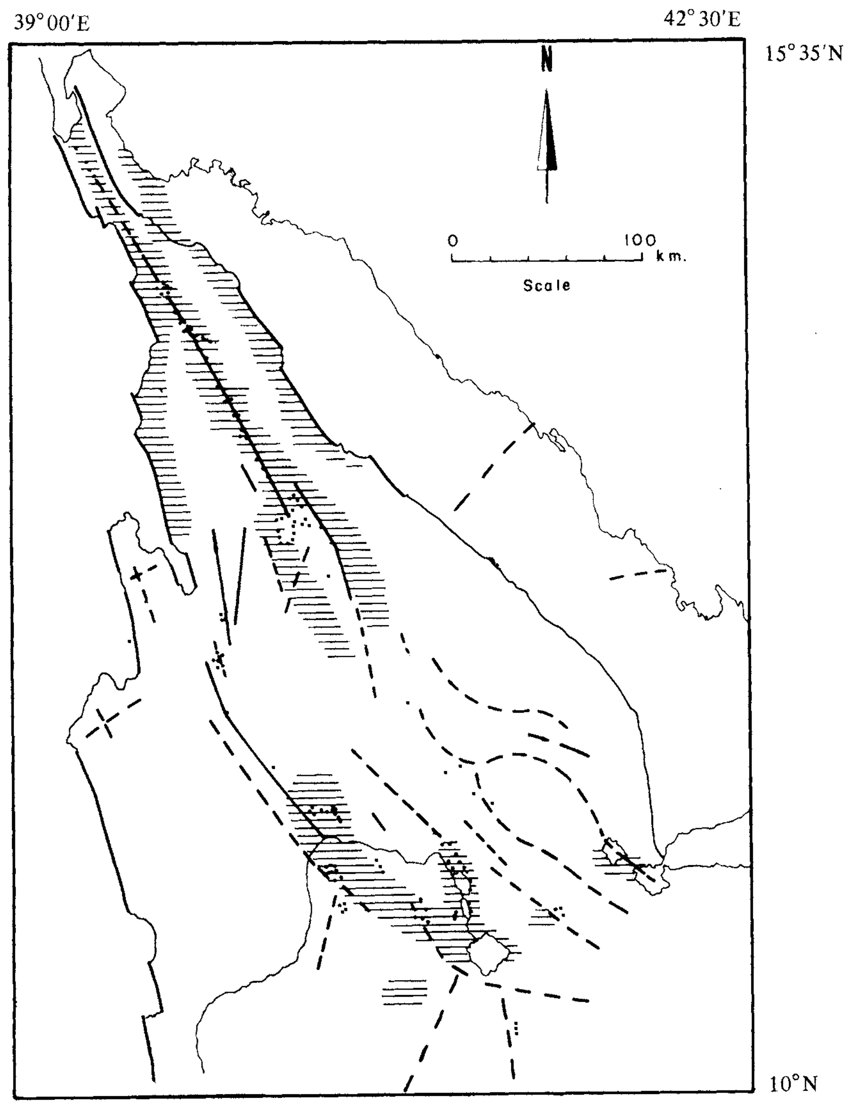

Fig.3. Sketchmap showing the distribution of present thermal manifestations (dots) and submarine volcanism (shaded areas) of the Afar depression, as well as a simplified picture of the main spreading axes (full lines: axial oceanic ranges; dashed lines: main graben axes).

Since the birth of the depression to the present time, volcanism and sedimentation of marine, lacustrine, evaporitic and continental facies have been simultaneously carried on (Faure et al., 1969; Bonatti et al., 1971). Long range correlations are therefore hazardous, and the use of such terms as "trap series", "Afar basalts", "Aden series" (Bannert et al., 1970; Mohr, 1971) is rather meaningless. All observed sequences in the Afar are of mere local significance and cannot be extrapolated or extended over large areas.

In the light of the structural pattern of Afar, three types of volcanism can be distinguished; these are considered in turn below.

Volcanism of oceanic nature

This type is located in the rift axes of the Red Sea–Gulf of Aden megastructure. It is best exemplified by the Erta'Ale range in northern Afar (Barberi and Varet, 1970), as well as by its southerly en-echelon continuations, Alayta and Tat'Ali ranges (Barberi et al., 1970). In central Afar, the volcanism of the N. Loggia range, of the Dama Ale centre, of the Manda range, and of the Asal graben, probably also belongs to this type. These volcanic units are characterized by a more or less advanced stage of magmato-volcanological evolution. From initial basaltic fissure eruptions it leads to shields of intermediate typical iron-rich terms, to strato- and even cumulo-volcanoes. The latter are of rhyolitic composition (alkaline to peralkaline), often with tectonically triggered local or extended rejuvenations.

The different axial ranges have a very similar petrology. Close relationships always appear between volcanological and petrological evolutions. The primary magma is of a transitional nature between tholeiites and alkali basalts. Fractionation of this magma has produced several intermediate types, leading to final peralkaline rhyolites. A complete differentiation series is found in the most complex units such as Erta'Ale and Tat'Ali ranges. Elsewhere, as in the Alayta range, differentiation has not proceeded beyond iron-rich intermediate ferro-basalts and dark trachytes. Everywhere, within these axial ranges, regular volume relationships characterize the proportions of various types of rocks in the differentiation series, with a highly marked predominance of basaltic lavas. This suggests a common origin for all these products, obtained through fractionation of basalts under low and decreasing oxygen fugacity (Treuil et al., 1971). The strontium isotopic values are remarkably low and uniform for all these lavas and point to a mantle origin without any sialic crust contamination (Barberi et al., 1970).

Continental volcanism

Along the rift margins, volcanic units are quite different from the axial ones. In this tectonic environment, central complex volcanoes are frequently affected by summit calderas, associated with acid lava domes and flows, ignimbrite sheets, pumice beds and basalt lava fields. Sialic products are always largely predominant as compared to basaltic ones, and intermediate lavas are scarce or lacking. Basalts are fissural, with their feeder vents obviously distinct from the central volcanoes. Chemical and petrological characteristics of these volcanic massifs show so marked a contrast between their acidic and basaltic constituents that a differentiation origin for the former seems unlikely (Fig.4). Moreover, isotopic ratios show wide variations from normal low values in basalts to high ones in sialic rocks; the latter are typical of crustal material. This would seem to indicate that either contamination or direct crustal melting has played a role in the genesis of these sialic rocks (Barberi et al., 1970).

Stratoid volcanism

Central and southern Afar volcanism consists mainly of extensive, thick, stratoid series of fissural basalts (traps) with associated ignimbrites and some silicic central piles. Their location seems to be generally connected with the intersection of the main trend with other fault directions, sometimes local ones as for instance around the Affara granitic batholith. Some of these volcanoes are quite large, for example Gad'Elu or Ayalu, which are 1000 m to 2000 m high. Others are much smaller and form plugs or domes apparently associated with ignimbrites. They may be considered as final products of ignimbritic eruptions, similar to the Novarupta stage of the Valley of Ten Thousand Smokes outburst (Bordet et al., 1963).

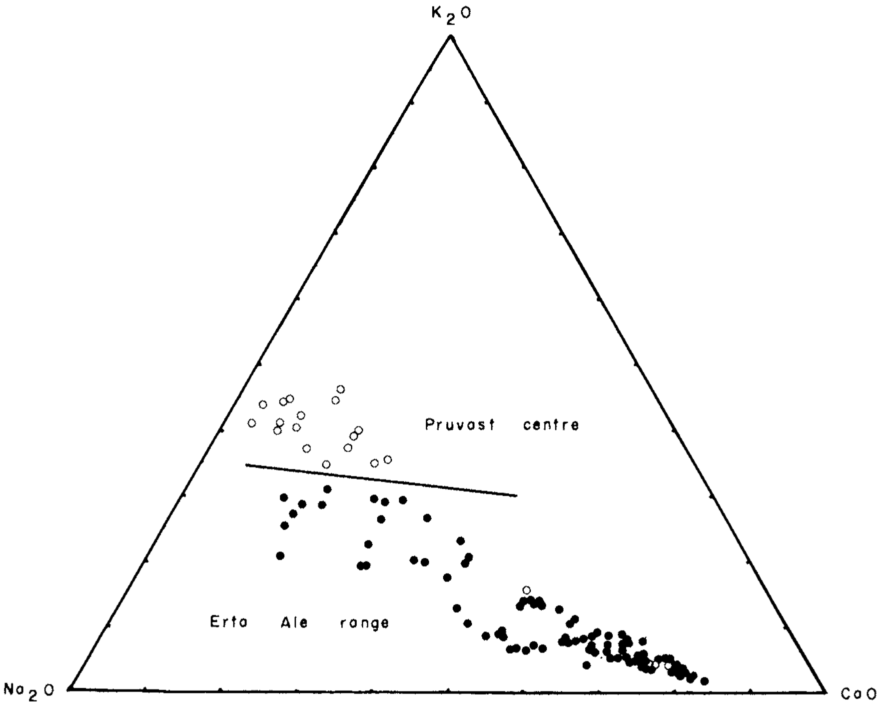

Fig.4. $CaO-Na_2O-K_2O$ diagram for axial and lateral volcanic rocks. A regular trend is observed in Erta'Ale range (axial), while potassium richer silicic rocks of Pruvost centre (lateral) are clearly outside the trend.

Comparatively young fault-scarps affecting this series exhibit vertical sections, up to 1000 m high, generally without any break or unconformity. The base of these accumulations remains hidden underground. K–Ar determinations show ages ranging from 0.4 to 11.1 m.y., but it must be borne in mind that this type of activity should have started far earlier (?25 m.y.), and that very fresh-looking surfaces of aa-type basalts or blister-affected ignimbrites occur frequently on the top of these series. Evidence indicates that these

stratoid volcanites have been emitted through fissures of the megastructure directions north from latitude 11°00′N, through fissures of Ethiopian direction south from this latitude.

The petrological sequence of the stratoid series and associated acidic centres, which are composed mainly of transitional basalts and rhyolites of mostly peralkaline composition, are similar to the sequences of the Ethiopian Rift. The origin of the fissural pantelleritic ignimbrites and of the centrally emitted lava-flows and ignimbrites have not been elucidated so far. The large volumes of these peralkaline volcanics, their association with basalts in fissural activity, and the scarcity, or absence, of intermediate rock types, seem to rule out any direct petrogenetic link between flood basalts and associated silicics, although both are in close tectonic association and transitional basalts are known to be potentially genetically related to pantellerites (Coombs, 1963; Barberi et al., 1971a). The very high soda content of these peralkaline rocks is difficult to explain by a mere fusion process of ordinary sialic crust. The large extent of such peralkaline products in rift systems may suggest the existence of some tectonic control for this type of volcanism, and their probable origin from lower crust or mantle depths seems indicated by their common fissural origin.

Granitic intrusions

Lastly, it should be stressed that several granitic intrusions outcrop in Afar. They are typically aegirine-arfvedsonite-cossyrite peralkaline granites, with well developed graphic textures. They correspond to the crystallization at depth of liquids of pantelleritic composition. Isotopic studies under progress should help to solve the problem of their relationships with the peralkaline rhyolites as well as the question of their origin, obviously related to the early stages of opening of the Danakil Rift (Barberi et al., 1972; Black et al., 1972).

Activity

Present-day fumarolic and eruptive activity (Fig.3) occurs along the axial volcanic ranges, in association with acidic central volcanoes or at the intersection of the main trends with other tectonic directions. This distribution may be related to two different heat release origins. High temperature manifestations within the axial ranges could be due to a rise of upper mantle material beneath (Marinelli, 1971). Permanent pockets of magma are hardly conceivable in areas characterized by fissural activity. This is rather similar to the situation along the crests of mid-oceanic ridges with their associated high heat-flow. The second type of surface heat manifestations could be related to the presence of magma pockets, located at the intersection of two tectonic directions or underneath existing silicic volcanic centres.

Volcanic activity is usually fumarolic, but eruptions happen comparatively frequently; at least three occurred during the last century (Alayta, Dubbi and Erta'Ale). Erta'Ale volcano is exceptional in having a sub-permanent lava lake within its pit crater. Discovered by Pastori in 1906, we have kept it under observation for four years (Varet, 1971).

CONCLUSIONS

Afar volcanism appears as a rather exceptional, tectonically controlled association of typical oceanic basaltic volcanism (with accompanying intermediate or silicic differentiates) and typical continental rift volcanism (characterized by trap series with large peralkaline silicic rocks).

The concept of an "Afar triple junction" should perhaps be reconsidered, since the evidence given here shows that the Red Sea and the Gulf of Aden constitute one and the same oceanic megastructure. The main Ethiopian Rift, on the contrary, is a continental structure and merely runs into the oceanic one. The so-called Afar triangle is therefore made up by only two main tectonic units, viz., Afar proper, i.e., the Red Sea–Gulf of Aden inflexion area, which occupies more than the 3/4 of the whole depression, and the northern end of the Great Rift Valley of East Africa, located in the southern corner of the Afar triangle.

ACKNOWLEDGEMENTS

Prof. G. Marinelli and Dr. G. Giglia from Pisa University have taken a preponderant part in the elaboration of the present paper.

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Detailed field mapping of Afar has been carried out in the period 1967–1971, and is partly summarized in the CNR-CNRS 1/500,000 scale geological map of the Danakil depression (Northern Afar, Ethiopia) prepared by Barberi et al. (1971b) with the collaboration of E. Bonatti, S. Borsi, J.L. Cheminée, H. Faure, G. Ferrara, M. Martini, and E. Chédeville.

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A PRELIMINARY INTERPRETATION OF THE GRAVITY FIELD OF AFAR, NORTHEAST ETHIOPIA

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ABSTRACT

Makris, J., Menzel, H. and Zimmermann, J., 1972. A preliminary interpretation of the gravity field of Afar, northeast Ethiopia. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 31–39.

All existing gravity data in Afar, northeast Ethiopia, were uniformly reduced to Bouguer values and were compiled into a map of 10 mgal isogals. Five crustal models crossing Afar E–W were calculated. The results obtained were as follows:

In south Afar the crust is most probably of the continental type, slightly attenuated. Along the Wonji fault belt a series of relative maxima is built as a continuation of the Ethiopian Rift in Afar. The crust at the eastern edge of the Ethiopian Plateau is approximately 30 km thick and is strongly attenuated under the Wonji fault belt which is underlain by material of low velocity and density. Towards the Aisha horst the crust increases again to a thickness of approximately 20 km. The trends of the field are closely connected to the directions of the escarpments.

In north Afar the crust is very strongly attenuated and partly oceanized. The trend of the gravity field is parallel to that of the southern Red Sea and strikes NNW–SSE. South of 13° N this trend changes to NW–SE and seems to be the continuation of the Gulf of Aden. The Danakil mountains separate the Danakil depression from the Red Sea. They consist of sialic crust, which has been rotated in an anticlockwise direction and most probably are isostatically uncompensated.

INTRODUCTION

A summary of all gravity data in Afar uniformly reduced to Bouguer values and compiled in a map of 10 mgal isogals is given. It is aimed to study the behavior of the gravity field in connection to the tectonic state and structure of the crust. The gravity field will be qualitatively discussed and five crustal models crossing Afar (Fig.1) from E–W will be presented. The parameters used for the two-dimensional calculations were constrained from seismic evidence and geological observations obtained at the adjacent areas.

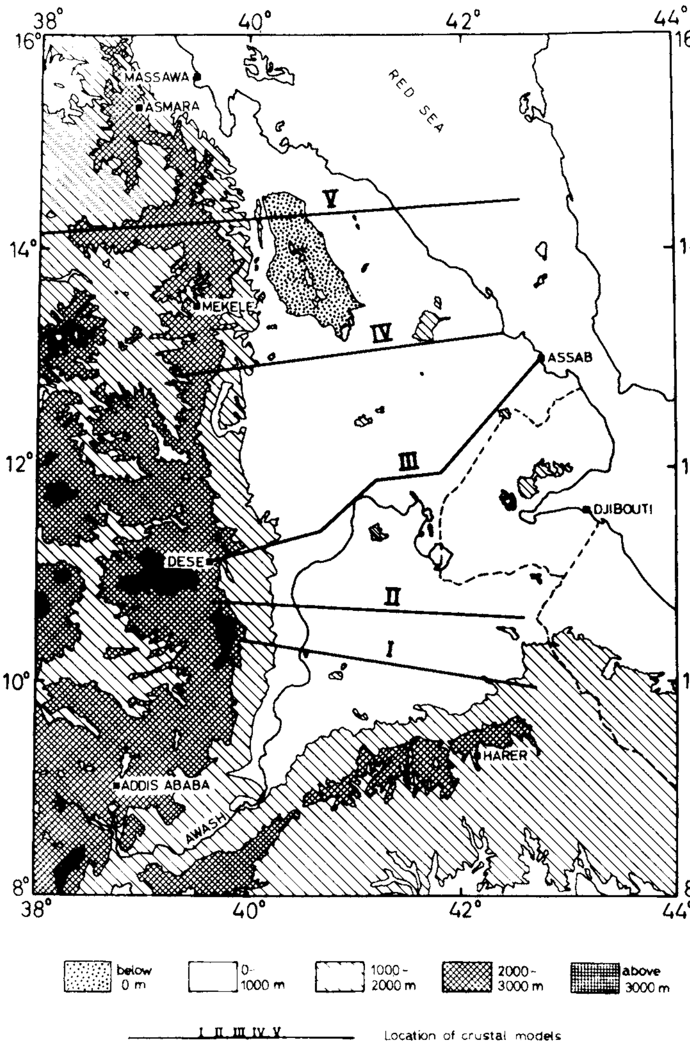

Fig.1. Location of the five crustal models crossing Afar in E–W direction.

COMPILATION OF THE BOUGUER MAP

The gravity data used for the compilation of the Bouguer map were obtained from different sources. North Afar has been mainly measured in 1968 by Prakla-Seismos GmbH on behalf of the Salzdetfurt AG, Hannover (Ochse and Ries, 1972). Gravity values of the Geophysical Observatory, Haile Sellassie I University, Mohr and Rogers (1966), Mohr and Gouin (1967), Gouin and Mohr (1964) were used as well.

South Afar was covered by approximately 1000 gravity measurements established in 1969 and 1970 by Makris et al. (1970, 1971). A total of approximately 1,800 stations

were uniformly reduced and compiled into a map of 10 mgal isogals. The mass reductions were spherical to Hayford zone O_2 with uniform density of 2.67 g/cm^3. The gravity data have been reduced to the mean sea level. The normal field was computed by the international formula of 1930.

The map was extended along the coast of the Red Sea to the Sudan border by data measured by Naftaplin in 1960 (Filjak et al., 1960) and into the southern Red Sea with measurements that C. Morelli (personal communication, 1971) put at our disposal. The Red Sea data have not been corrected for topographic effects.

A QUANTITATIVE INTERPRETATION OF THE BOUGUER MAP

The gravity field of Afar (Fig.2) at its southern and western limits is effected by the Somali and the Ethiopian Plateau respectively. The isogals follow closely the topographic features. There are, however, two distinct differences between the southern and the western limits:

(1) The gradient of the gravity field towards the western border (Ethiopian escarpment, 3.5–4.5 mgal/km) is steeper than towards the southern one (Somali escarpment, 1.5–2.5 mgal/km).

(2) The border of the Ethiopian Plateau towards the depression has a greater negative Bouguer anomaly (approximately –220 mgal) than that of the Somali Plateau (approximately –160 mgal).

Considering these two facts we must expect that the crust at the border of the Ethiopian Plateau is thicker than that of the Somali and that the tectonic distortion of the crust towards the depression is much stronger under the Ethiopian than under the Somali escarpment.

The gravity pattern of South Afar is strongly influenced by a swarm of local anomalies, extending from the Ethiopian Rift into the depression. The general trend of these anomalies is NNE, and they are distributed between Metahara and the Abbe and Gamori Lakes over a distance of approximately 400 km along the Wonji fault belt, Mohr (1967).

North of Lake Abbe the gravity field is aligned in a NNW–SSE direction. The continuation of this trend into French Somaliland seems to be probable. Gravity measurements in French Somaliland and the adjacent territory to the north could provide direct proof that the Gulf of Aden Rift coming through the Gulf of Tadjoura does not extend westwards as postulated by Girdler (1970), but turns NW and then NNW through north Afar, parallel to the trend of the Red Sea, as suggested by Tazieff (1971).

In north Afar the gravity field strikes NNW and the central trough of the Salt Plains – Danakil depression – has Bouguer values of between 0 and –20 mgal. This area seems to be an extension of the Red Sea into Afar. This fact is strengthened by model calculations which will be shown in detail in the following section and also by petrological studies. Barberi et al. (1970) proved that the active volcanic range of Erta'Ale is composed of subcrustal material uncontaminated by sialic crust.

The Danakil horst extends to the east of the Salt Plains along the coast of the Red Sea

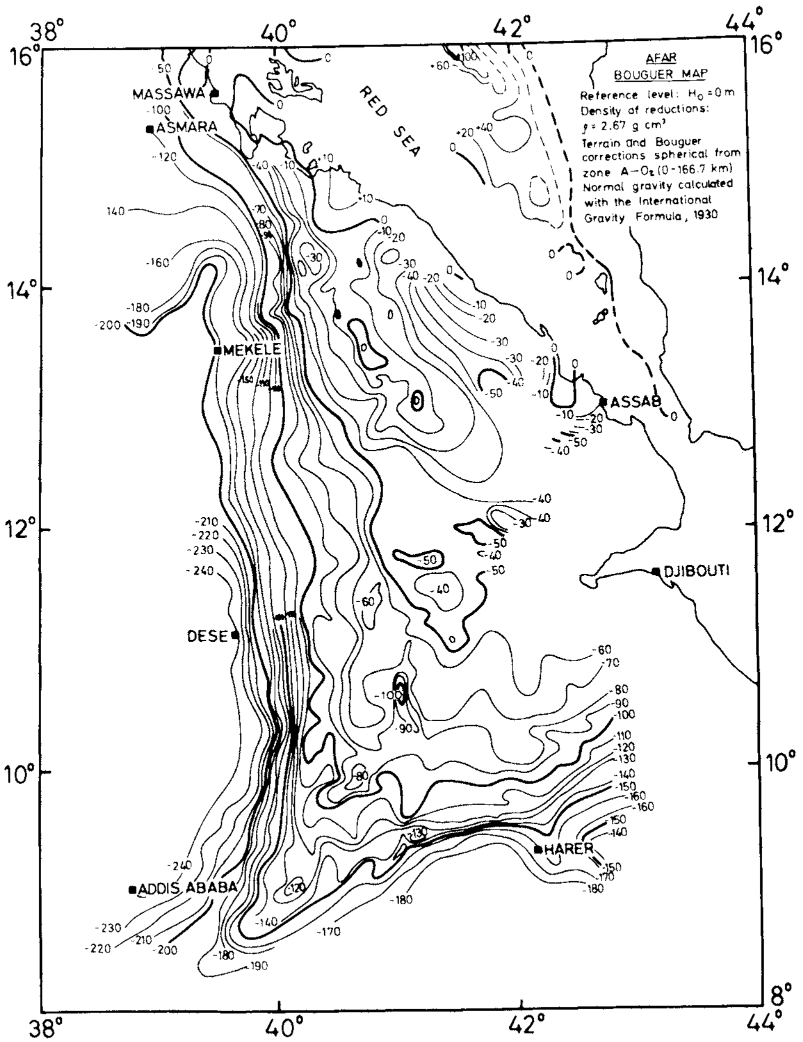

Fig.2. Bouguer map of northeastern Ethiopia.

and has maximum negative Bouguer values of approximately –40 mgal. The detailed distribution of the gravity field of these mountains is not sufficiently known. Further measurements are necessary in order to secure the extrapolated isogals. Along the Red Sea coast the gravity values are characterized by zero anomalies. Unfortunately the number of stations available is too small, and between Edd and Assab hardly any gravity measurements exist. The continuation of the NNE-trends north of the Lake Abbe are not gravimetrically confirmed.

CRUSTAL MODELS OF AFAR BASED ON GRAVITY AND SEISMIC DATA

Five crustal models were calculated across the Afar depression striking E–W (Fig.3–7). The crust consists of 2 strata lying on either low velocity upper mantle or rift material.

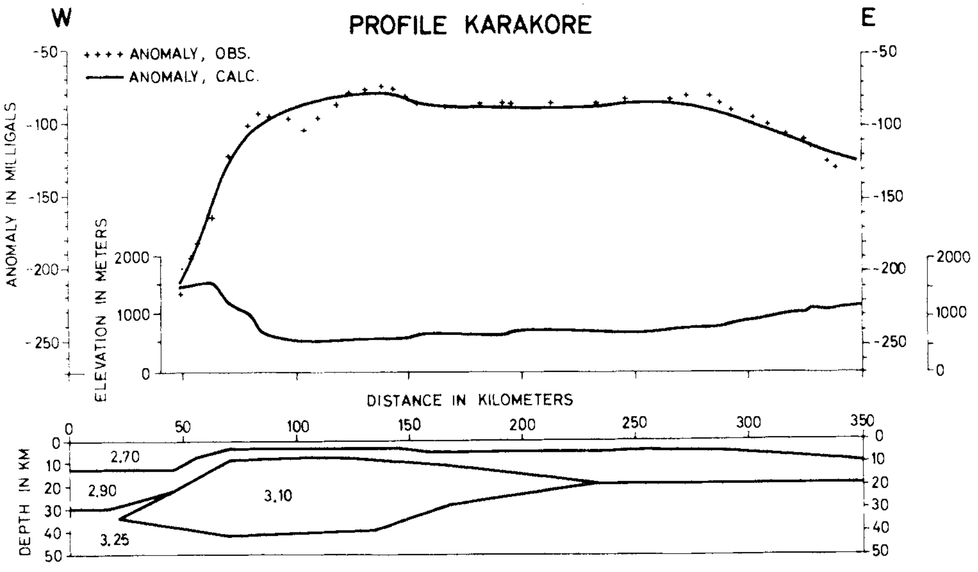

Fig.3. Model I: south Afar

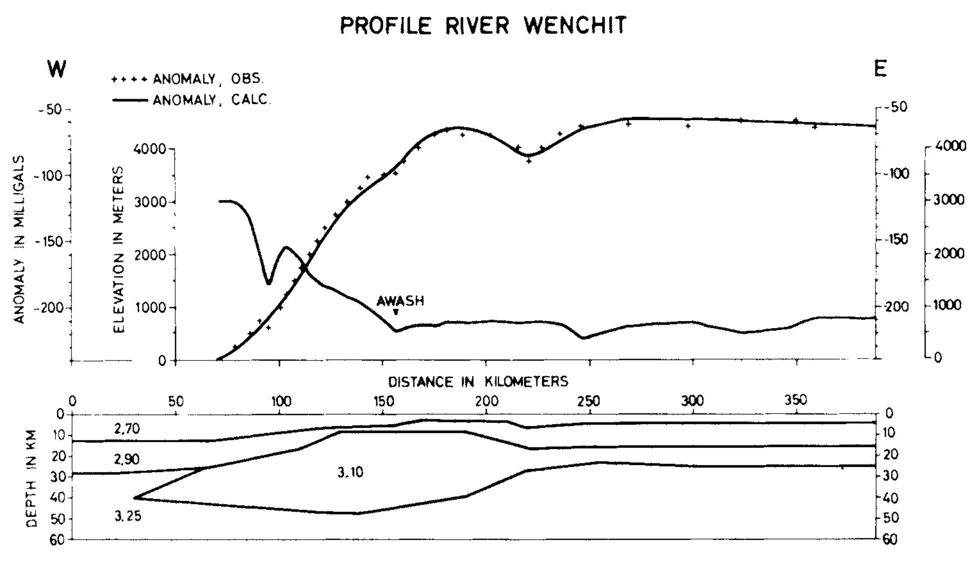

Fig.4. Model II; south Afar

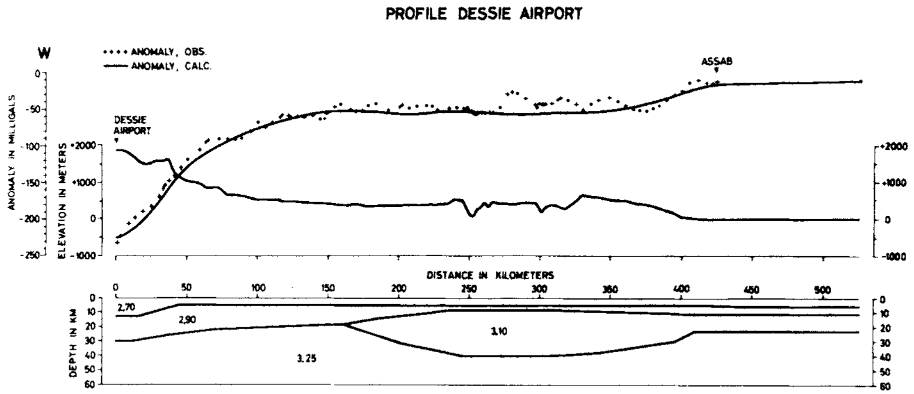

Fig.5. Model III: central Afar

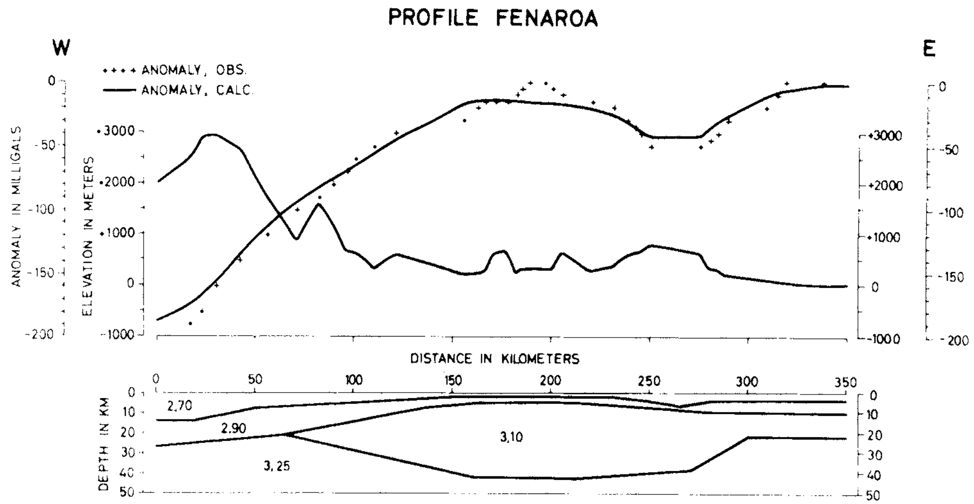

Fig.6. Model IV: north Afar

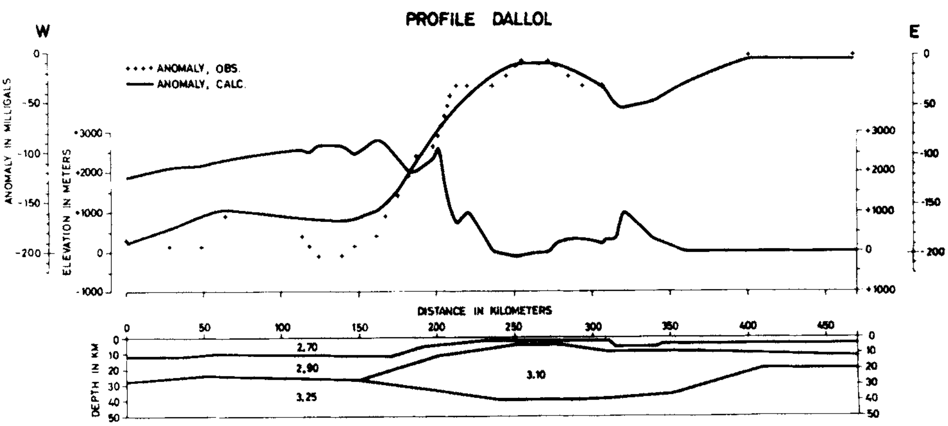

Fig.7. Model V: north Afar

The parameters used in these calculations are listed in Table I.

These parameters were chosen according to the latest data published on the African rift. Their reliability has greatly increased compared to that of previous studies, particularly since refractional seismic experiments (Griffiths et al., 1971), have revealed more precise P-wave velocities under the rift axis. The seismic studies of Bonjer et al. (1970) and Searle and Gouin (1971) have provided reliable data for Afar and the crust under Addis Ababa. Table II gives a summary of this information.

TABLE I

Parameters used in model calculations

Layer	Description	V (km/sec)	S.G.
1st	trap basalts, consolidated sediments, granites and gneisses	5.6	2.7
2nd	lower crust	6.6	2.9
3rd	intruded upper mantle differentiates or crust mantle mixtures	7.2	3.1
4th	upper mantle (lower density than normal)	7.95–8.0	3.25

TABLE II

P-wave velocities used in the 2-dimensional crustal calculations

	Upper crust (km/sec)	Lower crust (km/sec)	Rift material (km/sec)	Upper mantle (km/sec)
Bonjer et al. (1970)	–	6.0	7.2	–
Davies and Tramontini (1970)	3.99–4.55	6.11–6.86	–	–
Griffiths et al. (1971)	–	6.38	7.48	–
Laughton and Tramontini (1970) (Profile 6235/36, 6239, 6233)	3.98–4.33	6.15–6.54	7.06	8.16
Searle and Gouin (1971)	–	–	–	7.95

The comparison of the five models shows that Model I and Model II of south Afar (Fig.3 and 4) are of similar crustal structure. The steep gradients along the scarp of the Ethiopian Plateau are simulated by a rapid decrease in crustal thickness towards the depression.

The Plateau margin lies upon a crust 30 km thick which decreases to approximately 10 km at the rift axis. The crust thickness increases again towards the Aisha horst to approximately 20 km. The low velocity rift material (ρ = 3.1 g/cm^3) is located under the rift axis, close to the escarpment of the Ethiopian Plateau.

Model III between Dessie airport, Assab and the Red Sea (Fig.5) shows distinct differences compared to the first two models. The low velocity zone has moved far from the escarpment of the Ethiopian Plateau and has its maximum thickness under the annular

structures of the Dobbi, Guma and Gawa grabens. The crust thickness decreases gradually from 30 km under the Plateau to approximately 8 km under the graben.

The two traverses across north Afar (Fig.6 and 7) show steep gradients towards the Ethiopian Plateau and the Danakil Alps. In structure they are similar to Model III (Fig.5) differing only in the eastern part where the upper crust becomes thicker because of the Danakil Alps. The crust under the Danakil depression and the central volcanic ranges is very strongly attenuated. The 3.1 g/cm^3-layer rises to a depth of only 5 km under the central volcanic ranges. The crust towards the Ethiopian Plateau and the Danakil Alps is of continental structure. The Danakil mountains do not seem to be isostatically compensated.

CONCLUSIONS

From the distribution of the gravity field in Afar and from 5 crustal models which were computed using gravity and seismic information the following results have been obtained:

(1) In south Afar the crust is most probably of the continental type slightly attenuated. The field closely follows the Somali and Ethiopian escarpments and is partly disturbed by a swarm of maxima aligned NNE–SSW. They seem to follow the Wonji fault belt and are the continuation of the Ethiopian Rift towards the depression.

(2) In north Afar the crust is strongly attenuated and partly oceanized. The trend of the field is parallel to that of the Red Sea. The Danakil Alps between the Arabian and the Nubian plates are a sialic fragment which most probably is not isostatically compensated.

(3) The gradual change of the gravity field from a NNW–SSE to a NW–SE trend indicates the possibility that the Gulf of Aden continues into Afar in a NW–SE direction.

ACKNOWLEDGMENTS

The authors gratefully acknowledge the kind assistance of the Imperial Ethiopian Government, Ministry of Mines, Addis Ababa. Field measurements and evaluation of the data were made as part of a geoscientific programme in Afar sponsored by the German Research Association (Deutsche Forschungsgemeinschaft). Prof. C. Morelli has kindly put at our disposal gravity data of the Red Sea.

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A GRAVITY SURVEY OF THE CENTRAL PART OF THE ETHIOPIAN RIFT VALLEY

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ABSTRACT

Searle, R. and Gouin, P., 1972. A gravity survey of the central part of the Ethiopian Rift Valley. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 41–52.

SUMMARY

Gravity traverses have shown that the smooth negative gravity pattern over the Ethiopian and Somali Plateaus is disturbed by a strong positive anomaly over the entire width of the Ethiopian Rift Valley. On this broad positive anomaly are superimposed other smaller positive anomalies which were found to coincide with the location of the Wonji fault belt.

The present close-grid gravity survey of the central part of the rift defines more accurately the location of these positive anomalies. Correlations are made with the main geological and tectonic features of the rift and crustal models are presented which account for these anomalies.

INTRODUCTION

Reconnaissance gravity surveys carried out in Ethiopia by the Geophysical Observatory of Haile Sellassie I University since 1960 have revealed a broad negative Bouguer anomaly over the entire Ethiopian–Somalian swell (Gouin, 1970, fig.7 and 8). This anomaly has a minimum near (but not at) the central part of the Ethiopian Rift Valley, and is thought to be caused by low density upper mantle (possibly asthenosphere) underlying the Ethiopian highlands (Searle and Gouin, 1971, 1972).

Within the main Ethiopian Rift Valley, relative positive Bouguer anomalies are superimposed on this regional gravity low. There is a broad positive anomaly of about 100 km width and 50 to 60 mgal amplitude, which covers most of the rift, on which are locally superimposed much narrower positive anomalies. The latter have been interpreted as being due to intrusions associated with the Wonji fault belt (Gouin and Mohr, 1964). Gouin (1970) considered that the wider positive anomalies were not necessarily caused by high-density intrusions or severe crustal thinning. Searle (1970a), however, has shown that similar anomalies in the Kenya Rift Valley must be so interpreted.

This paper describes a detailed gravity survey of the central part of the Ethiopian Rift

Valley. This was undertaken to define accurately the positive anomalies, to compare them with those found in Kenya, and to see how closely the small gravity highs are correlated with the Wonji fault belt.

DATA

During 1970 and 1971, 420 new gravity stations were occupied in the area shown in Fig.1. Measurements were made at 2 to 5 km spacings along all extant roads and tracks in the area and along the shorelines of lakes. Some stations had already been established in this area by Gouin and Mohr (1964) and Mohr and Gouin (1967, 1968). Since the height control of these earlier measurements was generally less accurate, some of these stations were re-occupied during the present survey. Gravity values were measured using the Canadian Sharpe gravimeter no. 128 and the Lacoste-Romberg meter no. G-122; the probable errors are about 0.5 mgal.

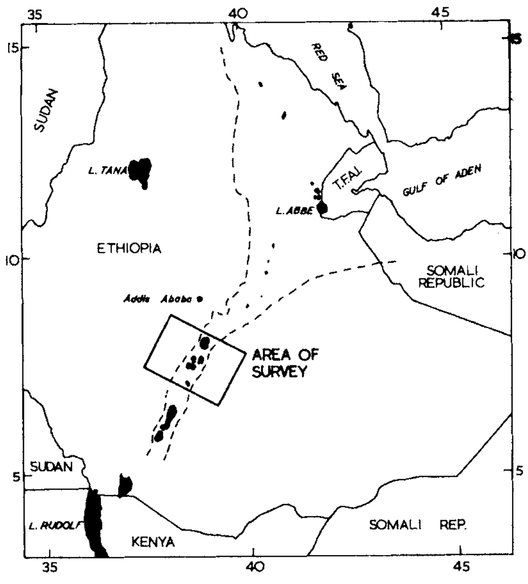
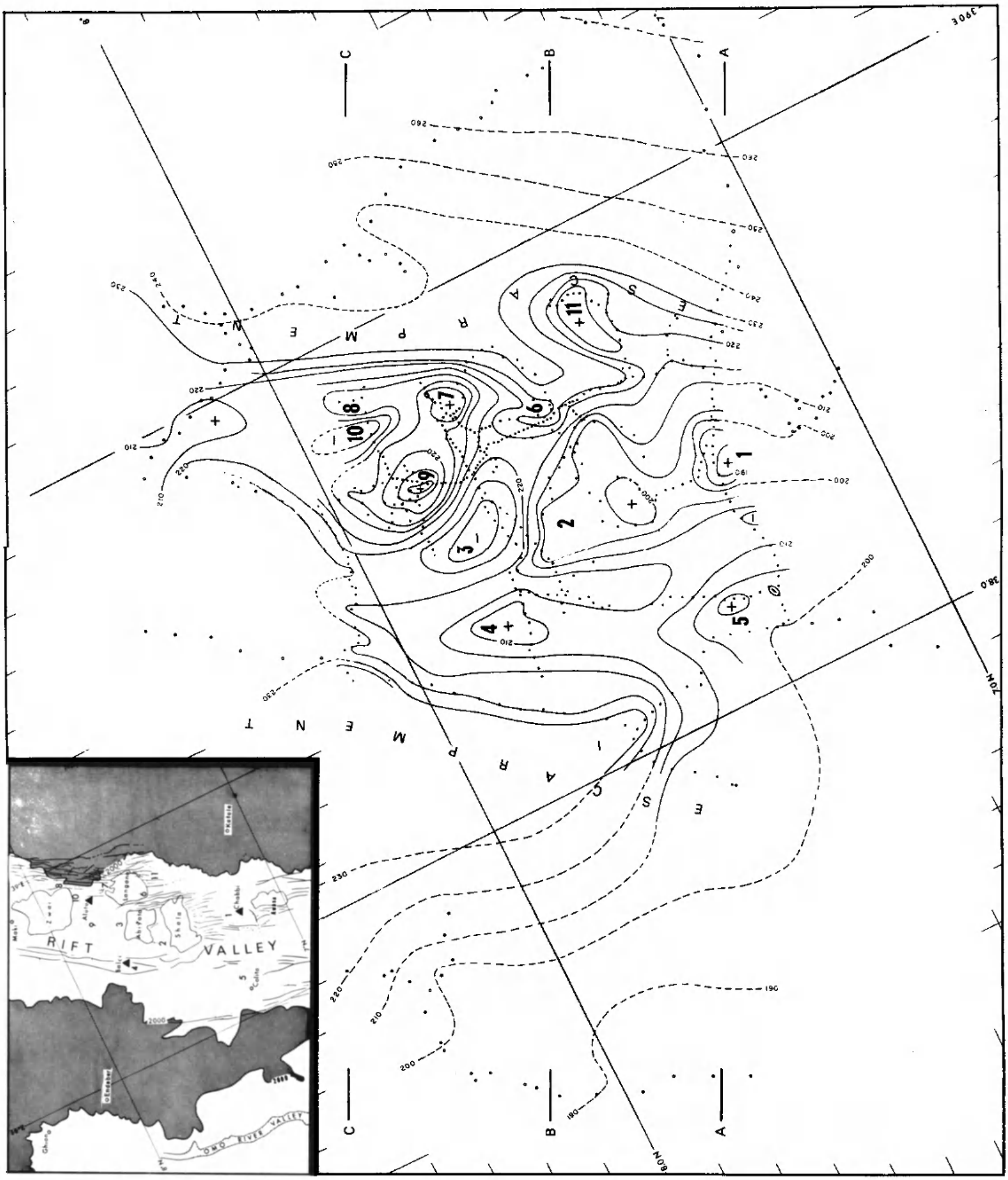


Fig.1. Location of survey area.

Fig.2. Main section: contour map of simple Bouguer anomalies, computed for a density of 2.67 g cm^{-3}, in the central part of the main Ethiopian Rift Valley. Full circles represent stations occupied or re-occupied during 1970–1971; open circles represent sites of earlier measurements. Full lines are isogals at 5 mgal intervals, based on the recent measurements, and broken lines are isogals at 10 mgal intervals interpolated from the older data. Inset: physiographic features for the area of the gravity map. Rift faults (mainly in the Wonji fault belt) are from Mohr (1972), this issue; silicic caldera volcanoes are shown by black triangles; land above 2000 m is shaded.
The numerals in the main figure and inset show the location of gravity anomalies described in the text.



RIFT
VALLEY
OMO RIVER VALLEY
Shala
Chabbi
ESCARPMENT
ESCARPMENT
C
B
A
C
B
A
1
2
3
4
5
6
7
8
9
10
11
8.0N
7.0N

Fig.2.

Elevations were measured using a single Paulin 7-inch Surveying Altimeter in "loops" of up to 2-h duration. Tests have shown that the elevations have a probable error of about 5 m. Since no small scale topographic maps of this area are available, station positions were plotted on uncontrolled aerial photo mosaics. Geographic coordinates were interpolated on the mosaics between about 20 points, whose coordinates were determined from 1:500,000 scale maps. The resultant coordinates of the gravity stations are therefore not accurate to better than about 0.05° (5.6 km), although over large parts of the area the relative positions are accurate to better than 1 km. Because of the lack of contoured topographic maps, no terrain corrections have been applied. However, very few measurements were made in areas of high relief, and errors due to terrain effects would probably not exceed two or three milligals.

Simple Bouguer anomalies were computed assuming a density of 2.67 g cm^{-3}. The total probable errors of the Bouguer anomaly values are about ± 2 mgal. Fig.2 shows a contour map of the simple Bouguer anomalies in the area of study. The earlier measurements performed by Gouin and Mohr on the shoulders of the rift have been included to show the regional field outside the rift. The larger probable error (5–10 mgals) of these earlier data does not substantially affect the shape of the regional field.

INTERPRETATION

Correlation with surface geology

Examination of Fig.2 reveals a very complex pattern in the Bouguer anomalies, in contrast to the rather simple pattern found in Kenya (Searle, 1970a, fig.3). The eastern side of the map shows an apparently greater complexity than the west, probably due to a denser station spacing. It is believed that these complexities arise both from the rather complex near-surface geology, and also from the existence of a number of different intrusive centres. Mohr (1967) has published an extensive account of the geology of the Ethiopian rift system, and this work is drawn on extensively in the following discussion. The anomalies discussed below are numbered, and can be located in Fig.2.

The positive anomaly (*1*) running north from 7.3° N 38.4° E, is clearly associated with intrusion and recent vulcanicity. The southward continuation of the line of this anomaly passes through the recent acidic volcano Chabbi, and the northern end of the anomaly (*2*) passes through the caldera-complex of Lake Shalla and the volcanoes to its north. The anomaly thus follows the Wonji fault belt here, as was pointed out by Gouin and Mohr (1964).

At its northern limit, this positive anomaly changes abruptly to a negative anomaly (*3*) immediately to the north of Lake Abiyata. The high gradients indicate a shallow origin for this negative anomaly, and it is probably due to recent sediments deposited when the lake level was higher than at present. Much of present Lake Abiyata lies in a faulted trough, and it is suggested that this trough continues north of the lake, where it is completely

filled with sediments. The negative anomaly has an amplitude of 17 mgal, and assuming a density contrast of -0.7 g cm^{-3} a maximum thickness of 580 m would be required for the sediments.

To the west of anomaly *3* is a broad positive anomaly (*4*). Although not well defined by our work, owing to the difficulty of access to this area, it appears to run the whole length of the western side of the rift floor in the region considered here. North of 7.7° N, it is associated with a well developed marginal graben in the rift floor. At 7.7° N, this anomaly passes through the acidic volcano Balchi, and to the south of this the anomaly follows a strong alignment of cinder cones. The positive anomaly (*5*) in the southwest corner of the map cannot be associated with any obvious surface features; it may or may not be connected with the marginal anomaly *4* farther north.

To the northeast of Lake Shala, a small positive anomaly (*6*) begins at the southern end of the Lake Langano and develops north eastwards, culminating in a strong anomaly (*7*) centered on a small caldera which manifests strong geothermal activity. A small positive anomaly (*8*) runs NNE from here along the Wonji fault belt but it is not well developed.

To the northwest of anomaly *7* is the strongest positive anomaly found in this region (*9*). There are a number of small volcanic cones near anomaly *9*, and its centre coincides with an exposed rhyolite plug. In spite of this, there is surprisingly little tectonic and volcanic activity associated with such a large gravity anomaly.

Due east of anomaly *9* is the silicic volcano Alutu lying on the Wonji fault belt. No measurements have been made on the volcano because of expected large terrain corrections, which at present cannot be accurately determined. However, the volcano appears to be associated with a small negative anomaly, which may be due either to a low density silicic magma chamber, or to low density effusives. Just north of this is a stronger negative anomaly (*10*), of amplitude -11 mgal, which seems to be due to sediments at the southern end of Lake Zwai. Assuming a density contrast of -0.7 g cm^{-3}, a maximum sediment thickness of 380 m would be required to account for the anomaly.

Finally, in the southeastern corner of the map there is a small positive anomaly (*11*). This anomaly may extend north eastwards along the rift margin, and be analogous to the anomaly on the western margin of the rift. However, the eastern anomaly is not associated with a marginal graben and is not related in any clear way to the geology.

It appears then, that the Wonji fault belt in this region is always associated with a positive Bouguer anomaly, although the anomaly is not always well developed. This anomaly is offset at the same place as the fault belt, but complications from other sources (e.g., sediments) prevent the offsets being clearly delineated. There are also other gravity highs, at least some of which are associated with intrusive zones other than that associated with the Wonji fault belt. In particular, there are two gravity highs along the rift margins.

Models

For the purpose of fitting models to the data, three profiles across the rift are considered. The lines of the profiles are marked *AA*, *BB*, and *CC* in Fig.2. Profiles of the Bouguer

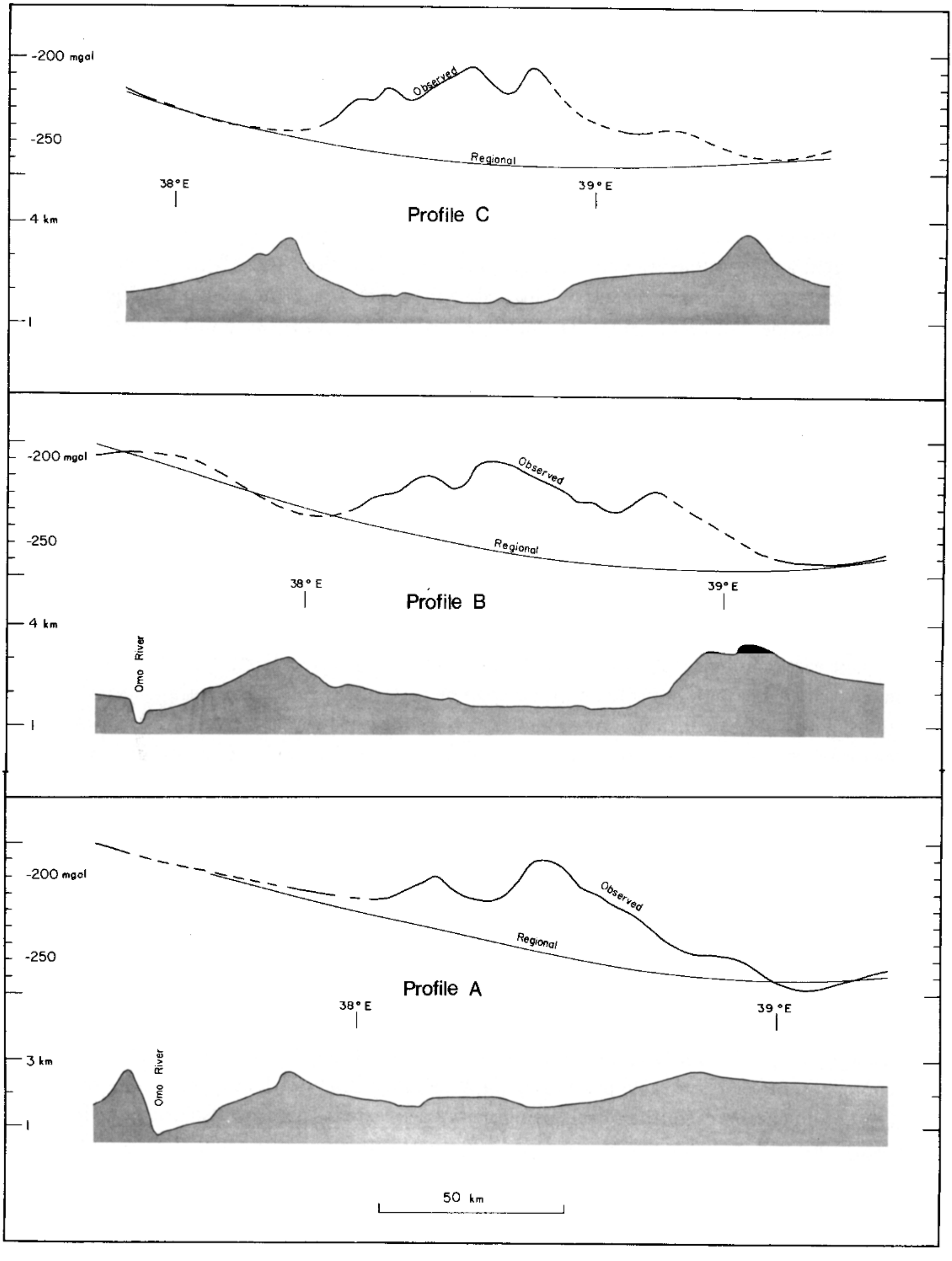


Fig.3. Profiles of Bouguer anomaly and elevation for the three profiles marked in Fig.2. The Bouguer profile has been taken from the contours of Fig.2, and is indicated by heavy continuous lines where there is a dense station spacing, and by broken lines elsewhere. The assumed regional anomalies used in the computations are shown by fine lines.

anomaly taken from the contours of Fig.2, and the corresponding elevation profiles, are shown in Fig.3.

It is clear that there is a relative positive Bouguer anomaly over the whole of the rift floor, and this anomaly is superimposed on the regional, broad negative anomaly of the Ethiopian plateaus. It can be seen from Fig.3 that the positive anomaly is inversely correlated with elevation. There is thus a possibility that the gravity high over the rift is merely due to too high a density having been used for the Bouguer correction. In profiles A and C the gradient of the Bouguer anomaly over the rift scarps can be reduced to zero if densities in the range 1.2 to 1.8 g cm^{-3} are employed for the Bouguer reduction. These values are considered to be unrealistically low; moreover, in profile B it is impossible to remove the Bouguer gradients assuming any density greater than zero. It is concluded that the positive anomaly is really due to a mass excess below the rift floor, although it is possible that its magnitude could be somewhat reduced if the true density of the surface rocks is in fact less than 2.67 g cm^{-3}. At present it is impossible to obtain a meaningful estimate of the average density of these rocks.

The regional anomaly was separated using the method described by Searle (1970a, p.20) to produce a smooth regional field in two dimensions. The regional field along the profiles is also indicated in Fig.3. It must be admitted that this regional anomaly is only approximate, the main uncertainties arising where the isogals of Fig.2 are interpolated between widely spaced measurements on the rift shoulders. Unfortunately, these critical areas are very difficult of access.

The error in the residual Bouguer anomalies arising from the approximate nature of the regional anomaly consists mainly of an uncertainty of perhaps 10 mgal in the datum of the residual anomaly, so it is still possible to produce meaningful, though necessarily imprecise, models from the data. The residual anomalies, with two sets of models computed from them, are presented in Fig.4 and 5. The gravity minimum near the centre of profile C is probably due mainly to low density lavas of the volcano Alutu, so no attempt has been made to fit the models to this feature.

Crustal and upper mantle models from gravity measurements across the Ethiopian Rift between 8° N and 9° N have been published by Makris et al. (1969, 1970). These models show a thinning of the crustal layers and an intrusion of upper mantle material beneath the rift. The models presented in Fig.4 of the present paper are based on similar concepts, although an attempt has been made to fit also the fine details of the gravity profiles. The density values given in the models are somewhat arbitrary, only the density contrasts between various bodies having been fixed in the computations.

The densities used for the crust in the models of Fig.4 are slightly higher than those used by Makris et al., although the same mantle density is used. There are thus smaller density contrasts between the upper mantle and lower crust, and also between the lower and upper crustal layers in the present models. In general this makes it easier to fit a model to a given profile, but in spite of this it was found rather difficult to fit the profiles given. It is certain that the amount of the anomaly generated at the crust–mantle interface cannot be greater than that given in these models (10–15 mgal). There is evidence of a

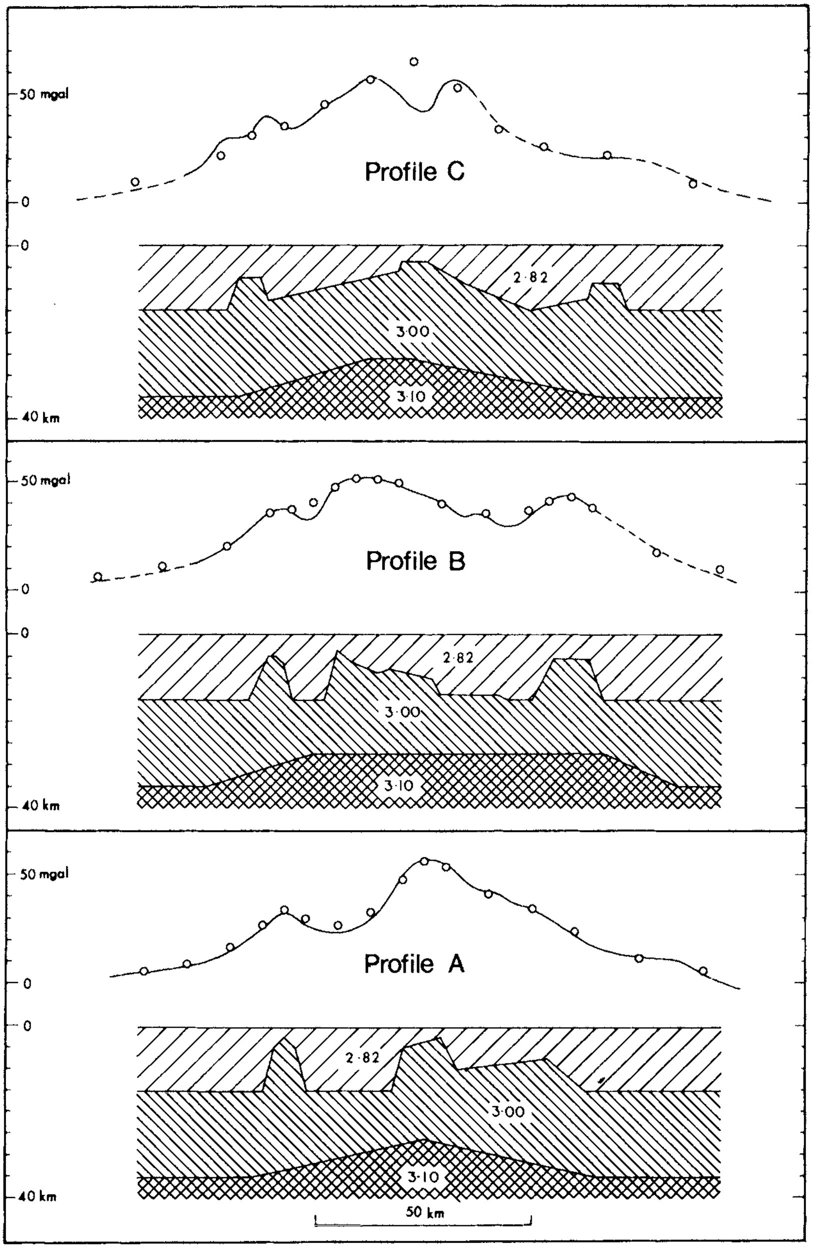


Fig.4. Residual Bouguer anomalies with models computed assuming a mass excess both in the upper and lower parts of the crust. Numbers on the models are densities in g cm^{-3}. Computed points are shown by circles. The gravity minimum near the centre of profile *C* has not been fitted for reasons given in the text.

layer with seismic velocity of about 7.4 km sec^{-1} under rift regions near the Gulf of Tadjoura (Lepine et al., 1972), in Afar (Searle and Gouin, 1972), and under the Kenya Rift (Griffiths, 1972; Griffiths et al., 1971; Long, 1972). This velocity would correspond to a density of 3.2 g cm^{-3}, i.e., higher than that used here in the gravity models. Therefore, if a similar layer exists under the Ethiopian Rift, the crust–mantle density contrast should be even greater than was assumed above. Consequently, any high-density body near the base of the crust cannot have a thickness of greater than about 5 km. This is not to say, of course, that no intrusion of mantle-derived material can exist in the lower crust; but if

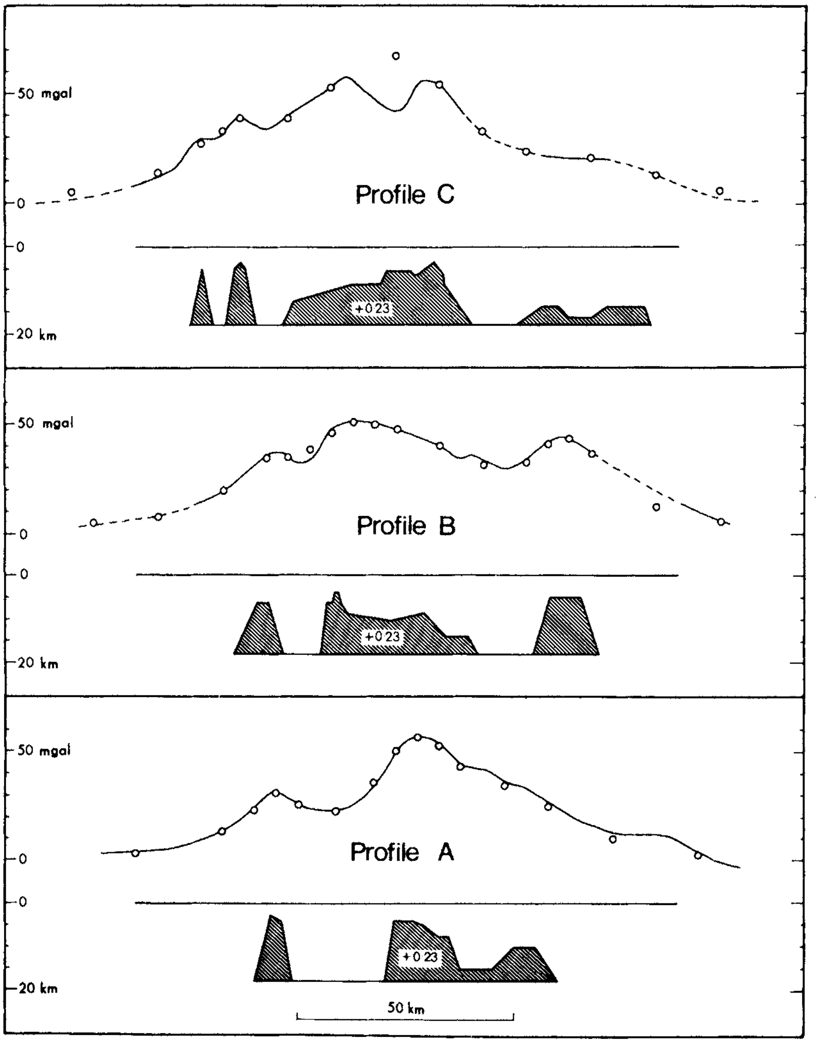


Fig.5. Residual Bouguer anomalies with models computed assuming a mass excess confined to the upper part of the crust. A constant density-contrast of +0.23 g cm^{-3} was used. The gravity minimum near the centre of profile *C* was not fitted for reasons given in the text.

it does (as seems likely), its density must not be much greater than that of the surrounding crust. This situation could arise by petrologic differentiation of the intrusive material.

Because of these considerations, models are also presented (Fig.5) in which the positive Bouguer anomaly is entirely explained by a positive density-contrast in the upper part of the crust. These models have the same constraint as model 2 of Searle (1970a) for the Kenya Rift, i.e., a base at 16 km below sea-level (18 km below surface) and a density-contrast of +0.23 g cm^{-3}.

The models indicate three major zones of intrusion. The central zone lies approximately along the axis of the rift floor, and the local peaks of this central zone correspond to the positions of the Wonji fault belt. As in Kenya, the models indicate a minimum depth of 3 km to the tops of the intrusions. In profile *A*, the peak of the central intrusive zone lies beneath the Wonji fault belt between Chabbi volcano and Shala caldera. On profile *B*, the main peak corresponds to the northern limit of the same section of the Wonji fault belt north of Lake Shala, but a subsidiary peak occurs to the east of this where the fault belt is dextrally offset and reappears at the southern end of Lake Langano. On profile *C*, the main peak is below the Wonji fault belt where it passes east of Lake Zwai, and there is a subsidiary peak just to the west of this beneath Alutu volcano.

Two subsidiary zones of intrusion are indicated by the models along the margins of the rift. A similar feature occurs in the Kenya Rift, but is much less marked. While it is true that the residual Bouguer anomalies are rather uncertain in the region of the rift margins, the magnitudes of these inferred intrusive zones seem too great to be entirely due to errors in the data. Moreover, at least the western intrusive zone of profiles *B* and *C* is clearly associated with volcanic and tectonic zones visible at the surface, as pointed out in the preceding section.

It is interesting to compare the width of intrusion in the three profiles. If one considers the width at a depth of 15 km, as fairly representative, it is found that the widths of intrusion for profiles *C*, *B*, and *A* are respectively 70, 55, and 40 km. If this rate of narrowing *is* continued southwards, the intrusion would die out completely about 120 km SSW of profile *A*, near Lake Abaya (6.5° N). Whether this actually occurs is unknown, but there is a positive gravity anomaly across the rift south of Lake Chamo, at about 5° N (Mohr and Gouin, 1968). Mohr (1967) has pointed out that the width of the rift decreases south of Lake Shala due to the influence of the Omo Valley immediately to the west of the main rift and whose tectonics are very similar to those of the main rift. Thus it is possible that some of the extension in this region may be taken up in the Omo Valley rather than in the main rift. Profile *C* lies beyond the northern limit of the Omo Valley tectonics, and should be uninfluenced by them. It is therefore considered that the best estimate of the extension across the rift in this part of Ethiopia is given by the width of the intrusion beneath profile *C*, i.e., 70 km. In this area, the trend of the rift (27° E of N) is almost perpendicular to the direction of opening of the rift, which can be estimated to be 114° from the rotation pole of McKenzie et al. (1970) or about 102° from the work of Mohr (1971). Therefore, in terms of plate tectonics, 70 km represents the extension parallel to the vector of relative motion between the Nubian and Somali plates, and this

is in good agreement with the prediction of McKenzie et al.

Searle (1970b) estimated the extension across the Kenya Rift to be about 30 km, but a different method was used; the extension was equated to the cross-sectional area of the intrusion divided by the depth of its base. If the extension in the Kenya Rift is redetermined using the same method as for Ethiopia, a value of 50 km parallel to the relative motion vector is obtained. The difference between this and the previous estimate of 30 km gives some idea of the imprecision of such devices to determine extension, due to the many uncertainties involved.

In spite of this, it is considered that a comparison between the extension, estimated on the same basis for the Kenyan and Ethiopian Rifts may be reasonably valid. When this comparison is made, it is found that the ratio of the extension in Ethiopia to that in Kenya is considerably less than the ratio predicted by McKenzie et al. It follows that the mean pole of rotation between the Nubian and Somali plates must be farther to the southwest than the pole of those authors, and this is in agreement with recent results obtained by Girdler and Darracott (1972).

CONCLUSION

The work described in this paper is in many ways only a preliminary study of the gravity of the Ethiopian Rift. Work is now in progress to extend the study northwards, to the southern corner of Afar at about 9° N. At the same time, G.R. Marsden of the University of Newcastle upon Tyne is making a gravity survey of the Rift between the area described in this paper and the Ethiopian–Kenyan border.

The preliminary results presented here confirm that the Wonji fault belt is, everywhere in the area studied, associated with a gravity high and an inferred zone of intrusion. However, other zones of intrusion also exist beneath the rift floor, particularly near the rift margins. The general nature of the intrusions beneath the rift in Ethiopia is similar to that in Kenya, although the amount of extension seems to be somewhat greater in the former case.

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AN AEROMAGNETIC SURVEY OF THE AFAR TRIANGLE OF ETHIOPIA

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(Received February 1, 1972)

ABSTRACT

Girdler, R.W. and Hall, S.A., 1972. An aeromagnetic survey of the Afar triangle of Ethiopia. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 53.

In 1968, an aeromagnetic survey was flown over the Afar triangle of Ethiopia. The survey extended from 9.5° N to 15° N and was bounded in the west by the Ethiopian scarp (about 40° E) and in the east extended over the Gulf of Tadjoura and the western part of the southern Red Sea. The flight height was 1.83 km (6000 ft). 34 profiles were flown in a direction N 100° and 42 profiles in a direction N 150°, the spacing between profiles being about 10 km. In addition, there were 14 tie lines. The total flight path was approximately 24000 km.

The magnetic and radioaltimeter records were digitised and profiles prepared showing both total intensity magnetic field (with I.G.R.F. epoch 1965.0 removed) and topography. Several analyses have been carried out: correlation studies were made in an effort to find trends and locate transform faults. Spectral analyses have yielded information on the shape, size and depth of the causative bodies and correlations between the magnetic anomalies and the geology are being studied.

In addition, a coloured magnetic anomaly contour map has been prepared with contour interval 100 nT, and scale 1:1,000,000 at 33° N. The contour map reveals several interesting features. First, the magnetic lineations previously mapped in the Gulf of Aden are found to extend westwards over parts of southern Afar. Secondly, there are regions of small amplitude, short wavelength anomalies which are thought to be associated with continental (sialic) material. Thirdly, the anomalies north of about 12° N have a NW–SE trend fanning towards the south.

The bearing of the survey on the tectonic evolution of the area is discussed. A full account is in preparation with large format diagrams.

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LOCAL EARTHQUAKE PHASES OBSERVED AT ADDIS ABABA, ETHIOPIA

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ABSTRACT

Searle, R. and Gouin, P., 1972. Local earthquake phases observed at Addis Ababa, Ethiopia. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 55–57.

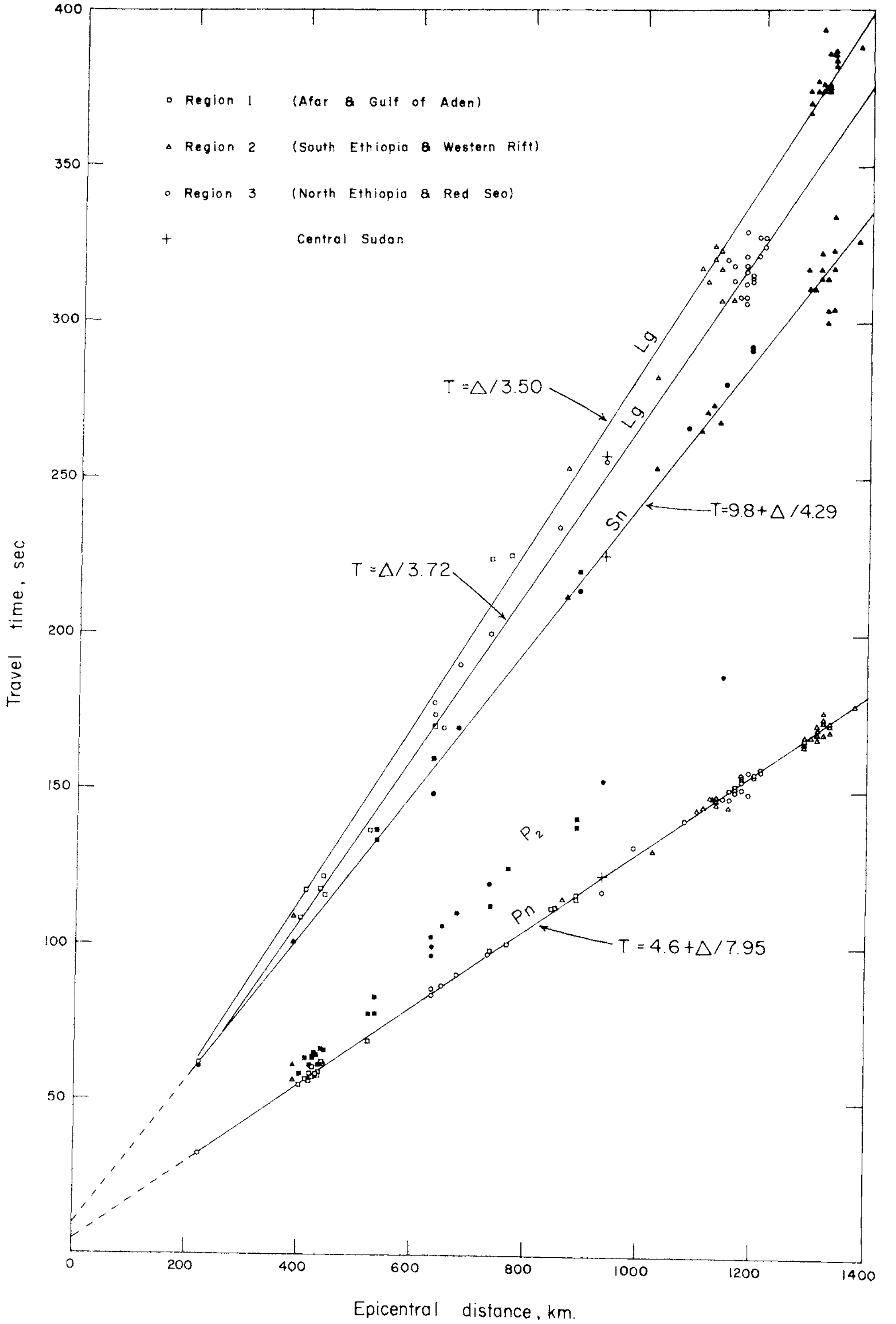
A summary is given of a work on local earthquake phases originating near the Afar triple junction, published in full by Searle and Gouin (1971).

SUMMARY

The records of 86 local earthquakes recorded at the World Wide Standard Seismograph station AAE, Addis Ababa, and occurring within 1400 km of the station have been analysed using the epicenter determinations of the National Oceanic and Atmospheric Administration Earthquake Data Reports. The phases Pn, possibly Pg and P*, Sn, and Lg were found. The epicenters were located mainly in three regions: the Red Sea, Gulf of Aden, and East African rifts (and at their junction in Afar). In addition, there were two isolated events in the Sudan.

Pn was observed on 80 of the records. Having analysed separately the events in all the different regions, it was found that only those from the East African rifts gave a statistically different Pn velocity. However, the travel time curve obtained from the latter events had an intercept which would imply a totally unrealistic crustal thickness under Addis Ababa; also, the data in this region were very poorly distributed with respect to epicentral distance, and therefore it is not considered that a regional variation of Pn in the area of this study can be proved at the time of writing. Taking all 80 events together, the average Pn velocity is 7.95 ± 0.03 (s.d.) km/sec. The Pn travel time curve (see Fig.1) has an intercept of 4.6 ± 0.5 sec, which places an upper limit on the crustal thickness under Addis Ababa of 48 ± 5 km.

A second P arrival could be seen on the records of most of the events within 930 km of the station, but not on the records of more distant events. There are insufficient data for travel time curves to be determined for this phase, or for the phase to be unambiguously identified. It is possible that both Pg and P* are being recorded. On Fig.1,


Region 1 (Afar & Gulf of Aden)
Region 2 (South Ethiopia & Western Rift)
Region 3 (North Ethiopia & Red Sea)
Central Sudan
Lg
Lg
Sn
P2
Pn
T =Δ/3.50
T =Δ/3.72
T=9.8+Δ/4.29
T = 4.6+Δ/7.95
Travel time, sec
Epicentral distance, km.
0
200
400
600
800
1000
1200
1400
50
100
150
200
250
300
350
400

these arrivals are marked P_2. However, Dakin et al. (1971) have shown that in the case of the 1969 Serdo earthquakes from central Afar, if (as seems likely) the second P arrival can be interpreted as Pg, and making the reasonable assumption that the mean crustal velocity along the propagation path is about 6.3 km/sec, there might be a basal crustal layer of velocity about 7.4 km/sec under western Afar.

Sn was recognised on the records of most events whose propagation paths did not cross rift valleys. No Sn was observed from events on the far side of the Red Sea or Gulf of Aden median rifts, or of the northernmost part of the Danakil depression in northern Afar, or of the Wonji fault belt south of 10° N. It was, however, transmitted across much of central Afar where, it is concluded, there are no major gaps in the lithosphere. The Sn velocity, based on 37 events, is 4.29 ± 0.04 km/sec, but the data are too few to test for regional variations.

Lg, with predominant periods of 1 – 2½ sec, and velocity 3.50–3.72 km sec^{-1}, was observed from most events, including many in the median rift of the Red Sea. Since in the latter cases the phase has travelled up to 290 km beneath the sea with little attenuation, it follows that not all of the Red Sea is floored by pure oceanic crust, and hence more severe restrictions apply to the opening of the Red Sea than are usually considered. On the other hand, observations of Lg from different locations in the Gulf of Aden are in agreement with the accepted distribution of oceanic crust there.

Finally, the Pn and Sn velocities lead to a Poisson's ratio of 0.29 for this region. Both the Pn and the Sn velocities found here are lower than those which have been reported for other parts of Africa, and the Poisson's ratio is considerably higher than normal. All this implies that the upper mantle here is of low density, probably has higher than normal temperature for a given depth, and is rather plastic. This is consistent with the raised topography and low gravity anomalies of the Afro-Arabian dome, and with the idea of a severely thinned lithosphere beneath the rift regions. Since the same physical properties have been found to extend at least as far as the central Sudan, we picture the Afar triple-junction as underlain by a body of asthenospheric material, wedging out away from the rifts to considerable distances.

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Fig.1. Plot of arrival time versus epicentral distance for the phases described in the text. Straight lines are the travel time curves by least squares analysis. P_2 and Sn are indicated by solid symbols to differentiate them from Pn and Lg. Based on fig.2 of Searle and Gouin (1971).

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SEISMIC PROFILES IN THE DJIBOUTI AREA

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ABSTRACT

Lepine, J.C., Ruegg, J.C. and Steinmetz, L., 1972. Seismic profiles in the Djibouti area. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 59–64.

The preliminary results obtained from the first seismic refraction profiles on the French Territory of the Afars and the Issas are presented. They can be summarized by a mean seismic section of the region centered on the Gulf of Tadjoura. It seems that the part of the Gulf of Aden characterized by an anomalously low mantle velocity, extends towards the west. The shallowing of the sea seems to be accompanied by the thickening of the crust. The section that was obtained shows a strong similarity to the Iceland section published by Palmasson (1971).

INTRODUCTION

The interpretation of all geophysical data obtained in the Gulf of Aden and the Red Sea supports the assumption that recent history of this area may be regarded as being mainly caused by the relative motions of three plates formed by the breaking of an initial shield (McKenzie et al., 1970). According to this general scheme McKenzie et al. consider that the Afar region should be mainly made of mantle material differentiated during the separation of Africa and Arabia.

In a detailed analysis of the regional structure of Afar, Mohr (1970) agrees with this global statement, pointing out however certain complementary peculiarities.

Among the problems still open, some may be studied by explosion seismology, namely, what structural condition is related with the upheaval of the region to sea level? where are the possible areas of deep injections? what is the extent of the remaining continental blocks outcropping in the horsts which are of great importance in Mohr's synthesis?

During March and April 1971 a first experiment of refraction seismology restricted to the Territoire Français des Afars et des Issas (TFAI) was carried out. It was completed by seismic reflection profiling with Flexotir in the Gulf of Tadjoura and Bab el Mandeb strait. Results of refraction profiles will be presented and discussed here.

DESCRIPTION OF THE EXPERIMENT

The geographical location of profiles, particularly of reversed profiles, was mainly limited to tracks and suitable locations where shots could be fired.

Two long profiles were chosen, one in an E–W direction with recording stations between Djibouti and Lake Abbe and shot points in the sea to the east, to connect with the work of Laughton (1969) in the Gulf of Aden; the other in N–S direction between Obock and Doumeira on the coast with shots east of the Gulf of Tadjoura and in the Red Sea, north of Bab el Mandeb. Recordings were made along four additional smaller profiles north of the Gulf of Tadjoura (Fig.1).

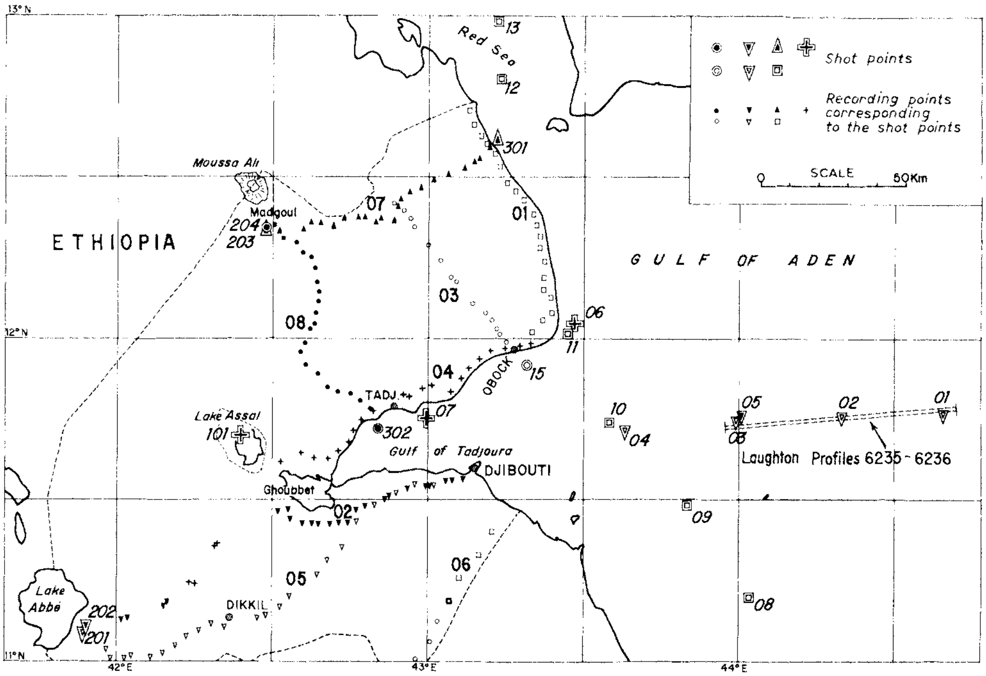

Fig.1. Location of seismic profiles carried out in 1971 in the Djibouti area. The recording stations and the corresponding shot points are represented by similar signs.

These multiple profiles were laid out with a view to obtaining a representative sampling of the structure in different regions and to examine the possible southern extension of the Danakil horst. The work of Tramontini et al. (1969) in the Red Sea shows that several profiles are necessary in such possibly heterogeneous regions to advance any valid interpretation.

Fifteen 180 kg shots were fired at sea by dropping bombs from an airplane. Shot depth was 40 m under sea level. Shot time was recorded by nearby telemetric buoys equipped with hydrophones. Shot locations were determined by the complex navigation system of

the plane. This shooting method, which may be improved, gave errors of 0.1–0.2 sec on shot times and 1/4–3/4 nautical miles for locations. The reduced accuracy in shot parameters is however sufficient in long range crustal refraction studies. Three other 180 kg shots were fired by a ship near the coast and four 300–400 kg borehole shots near Lake Abbe and in the Madgoul (north of the territory) were made to reverse some of the profiles.

To get travel time curves over a great distance range shots at sea were fired at increasing distances in the prolongation of the Djibouti–Lake Abbe and Obock–Doumera recording lines.

Each shot on a profile was recorded by twenty to thirty frequency modulated tape recording stations with 3-component, 2 Hz geophones.

Time base for the experiment was especially emitted by Radio Djibouti.

RESULTS

In spite of the rough climatic conditions and land configuration, the quality of records, although variable, is quite good. High wind noise level or anomalous conditions of wave propagation have disturbed the recordings in some cases.

In some instances the first stations are too far from the shot point to determine accurately the surface velocity, the values given are possible upper limits in the first layer.

The junction of travel time curves obtained on a recording line for shots of increasing distance is difficult because of the complexity of structures and reduced accuracy of shot times.

Preliminary results are summarized in two tables, the first for data in the distance range 0 to 100 km, the second for greater distances. Apparent velocities are derived by least squares and given to the nearest 0.1 km/sec. The intercept time errors are estimated between 0.05 to 0.2 sec for the shots in the sea, and to 0.02 sec for the terrestrial shots.

The estimated errors on the depth of discontinuities are therefore of the order of 0.2 to 0.5 km.

Three different velocity ranges can be seen in Table I: velocities less than 4.3 km/sec between 5.3 and 5.7 km/sec and between 6.3 and 6.7 km/sec.

We consider these velocities as characterising three different layers. Thicknesses of these layers and depth to the underlaying medium given in Table I are computed in assuming a horizontal layering beneath each shot point. The velocity of 5.5 km/sec is not found in all cases, which does not exclude the presence of the corresponding layer; the refracted wave being a first arrival only in a very small distance range may not be visible. The range of variation of thicknesses and velocities from one profile to another is relatively small. Possible local differences in structure will not be discussed.

The mean crustal section is:

Velocity (km/sec)	Thickness (km)
4.3	2.5
5.5	2.5
6.5	

TABLE I

Crustal structure in the Djibouti area according to the various profiles with velocity and thickness of first layers and depth of the lower medium

Profile number	Shot number	V_1 (km/sec) ± 0.1 km/sec	e_1 (km) ± 0.5 km	V_2 (km/sec) ± 0.1 km/sec	e_2 (km) ± 0.5 km	V_3 (km/sec) ± 0.1 km/sec
01	11	3.7	2.5	5.3	2.7	6.7
01	12	4.3	3.6	5.4	2.8	6.7
01	BC	–	–	5.3	–	6.5
05	201	4.2	3.7	–	–	6.4
03	15	4.2	4.0	–	–	6.5
04	06	4.3	2.5	5.7	2.5	6.7
04	07	–	–	–	–	6.3
07	301	4.3	3.6	–	–	6.5
07	203	4.2	3.7	–	–	6.3
08	302	–	–	5.7	4.3	6.3

TABLE II

Maximum observed apparent velocities at great distance

Profile number	Shot number	Apparent velocities (maximum)	Distance (maximum)
01	10	6.8	102
01	9	7.0	127
01	8	7.1	157
05	4	6.7	190
05	3	7.0	235
05	2	7.4	270
05	1	7.5	304

Table II gives the maximum first arrival apparent velocity observed at large distances for each shot, the profiles at these distances being unreversed. The apparent homogeneity of upper layers as revealed by the data limits to some extent possible ambiguous interpretations. The travel time curves do not show any clear breaking point suggesting the absence of any noticeable interface under the 6.5 km/sec layer. The velocity however seems to increase progressively. It is possible that a velocity contrast of few hundred meters per second is masked.

The travel time curve on profile Obock–Doumera has been inverted by the Herglotz–Wiechert method. The velocity of 7 km/sec is reached at a depth of 13 km. On the profile Djibouti–Lake Abbe the velocity of 7.5 km/sec is found at 35 km depth.

In summary, a preliminary study of our data gives the following structure: under a 5 km thick covering where the velocity reaches 5.5 km/sec a clear interface is found with 6.5 km/sec velocity beneath; velocity increases then progressively with the depth reaching values of 7 km/sec at 13 km depth and 7.5 km/sec at 35 km/sec (Fig.2).

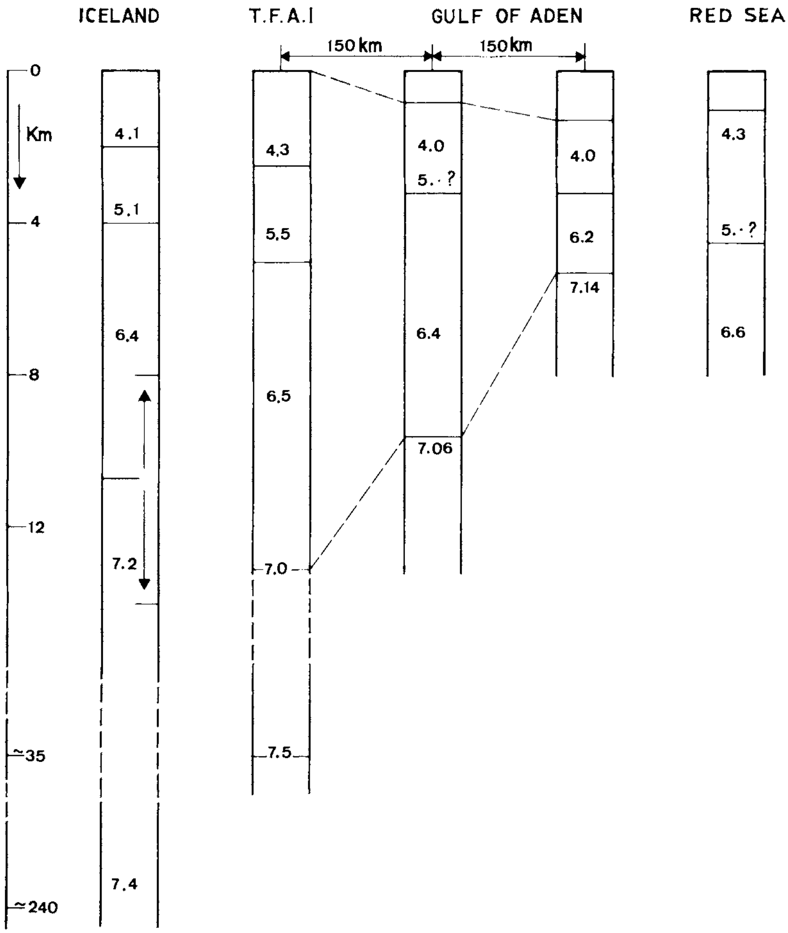

Fig.2. Comparison of crustal structure of the Djibouti area with neighbouring areas (Gulf of Aden and Red Sea) and Iceland.

INTERPRETATION

The above mentioned velocity-depth relation is valid for the area surrounding the Gulf of Tadjoura. The relation of the central zone of the Gulf of Aden with this region has been shown by the continuity of the axial trench in the Gulf of Tadjoura (Roberts, 1968), by studies on the seismicity (Fairhead and Girdler, 1970) and the pattern of magnetic anomalies (Girdler, 1970). One feature however varies rapidly in the E–W direction: the mean level of sea bottom is uplifted by 2000 m within 250 km. We consider that a continuity exists between the structure derived from refraction profiles in the axial zone of the Gulf of Aden and the structure we obtain in TFAI, the differences observed being causes or consequences of the relative uplift to the west. This opinion is supported by the fact that the differences are more important in the layer thickness than in the velocity.

Important regional upheavals are generally thought to be related to a mantle of anomalously low density and velocity. It must be pointed out that in our case the uplifting of the earth's surface to the west would be due to the combination of two features: increase in the thickness of overburden and accentuation of anomalous character of the mantle to the west where the crust–mantle boundary is no longer defined.

Fig.2 shows the velocity–depth sections derived from the present study compared with that derived by Palmasson (1971) for Iceland.

The structure found in the Gulf of Tadjoura area in continuity to that of the Gulf of Aden may be to some extend extrapolated to the north of the axial region by profile 01 west of Bab-el-Mandeb.

No confirmation as to a southern extension of the Danakil horst in the TFAI could be obtained from our data.

CONCLUSION

The preliminary results presented above give the following tentative conclusions: the mean velocity–depth distribution inferred from refraction profiles in the TFAI gives support to the concept of a continuity to the west of the axial zone of the Gulf of Aden and shows the accentuation of an anomalous upper mantle related to the general uplift of the area. This area may be extended in the northern part of the Gulf of Tadjoura.

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CONTINENTAL RIFT ZONES: THEIR ARRANGEMENT AND DEVELOPMENT

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ABSTRACT

Milanovsky, E.E., 1972. Continental rift zones: their arrangement and development. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics,* 15 (1/2): 65–70.

The rift zones may be classified as oceanic, continental and intercontinental (e.g., the Red Sea and Gulf of Aden). Most of the rift zones are connected and form the world rift system. A notable exception is the Lake Baikal Rift.

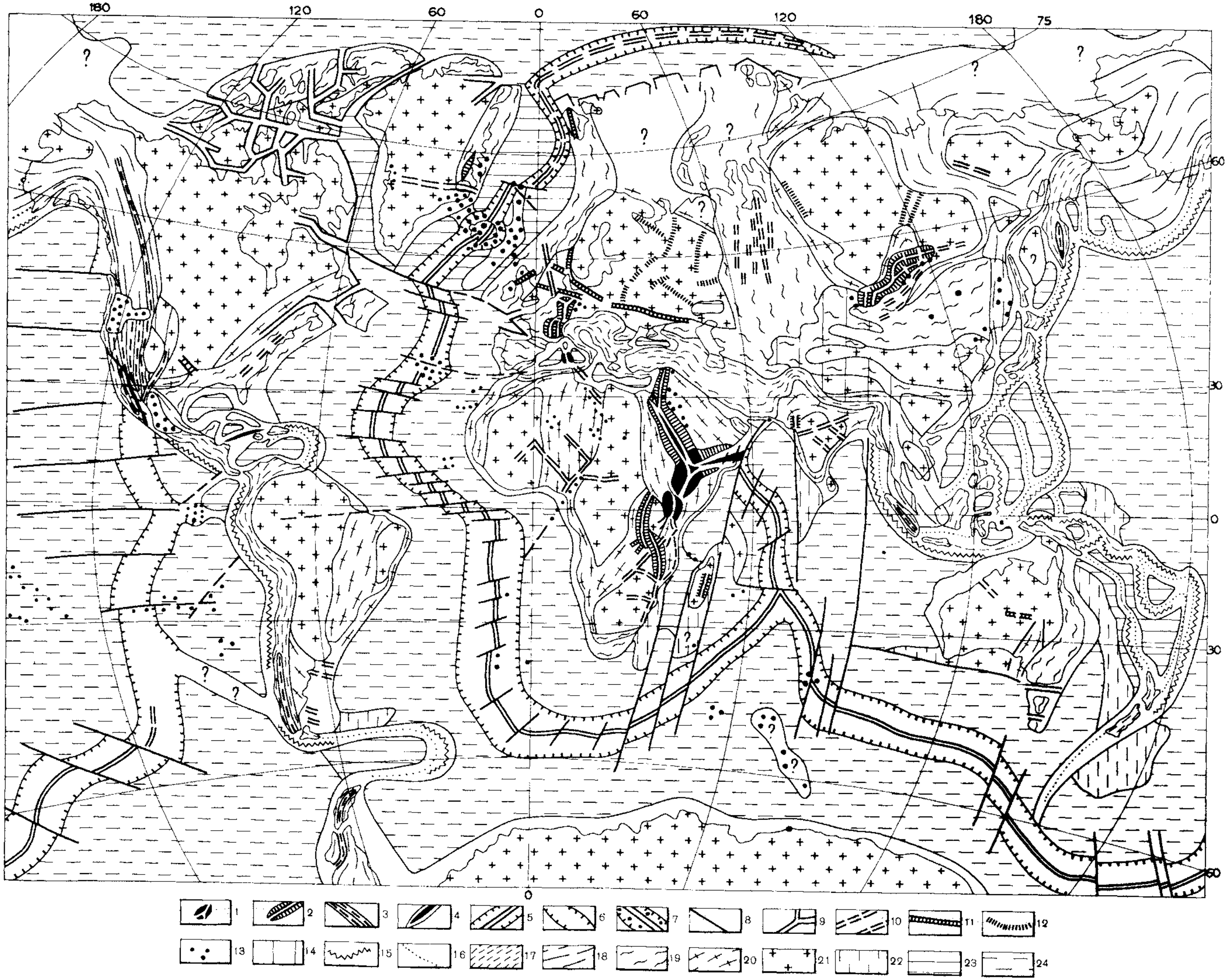
The Cenozoic continental rifts form two groups; the first associated with platforms and the second with young fold structures. The platform rifts may be divided into two sub groups, those associated with alkaline volcanism and updoming and those associated with little or no volcanism and minor up-doming. Volcanism in the rift belts of young fold structures is of calc-alkaline character.

INTRODUCTION

The rift zones are regions of stretching where the crust is somewhat thinned and the mantle has lower than normal density. The most distinctive form of rift is a relatively narrow, often step faulted graben with deep normal faults and uplifted margins. The rift zones are often characterised by high seismicity, magnetic anomaly bands, high heat flow and volcanism. In recent years, it has been recognised that the rifts are of worldwide continuity and complementary to the worldwide geosynclinal belts. Various stages of rifting may be recognised on the continents and in the ocean floor. This makes it possible to recognise several types of rift zones in a similar way to the various types of geosynclinal zones (Milanovsky, 1970 and Fig. 1).

MAIN CATEGORIES OF RIFT ZONES

There are three fundamental categories: (1) oceanic (intraoceanic) where the axial graben is bordered by oceanic crust; (2) continental (intracontinental) where both rift floor and shoulders are of continental crust (usually slightly thinned); and (3) intercontinental where the rift has oceanic crust and the shoulders have continental crust (examples include the Red Sea, Gulfs of Aden and California). The last may be supposed to be early stages in the formation of oceanic crust (Girdler, 1965). Thus these three categories reflect various stages in rift genesis.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MAIN TYPES OF CONTINENTAL RIFT ZONES

These may be divided into: (1) rift zones of platforms; (2) rift zones of young folded regions. The former are characterised by single axial grabens with alkaline volcanism and frequent carbonatites and the latter with numerous grabens and horsts with calc-alkaline volcanism.

Rift zones of platforms

These are associated with basement rocks which have a long, complex and varied history. The young rift structures often inherit the trend of the old basement structures or adjust to them producing geniculate, zig-zag or en-echelon fault structures. They may be divided into two:

Rift zones of arch-volcanic type

These are characterised by exceptionally intense and long volcanic activity such as the rifts in Ethiopia and Kenya (Milanovsky, 1969; Baker and Wohlenberg, 1971; Mohr, 1968). The igneous rocks are basic and intermediate lavas of a highly alkaline series. The volcanism is associated with uplift and arching and a 1–2 km deep graben develops along the axial part of the arch, often with branch rifts. The arching is associated with a Bouguer gravity minimum apparently due to melting in the lower part of the crust and top part of the mantle. The axial grabens are associated with narrow zones of gravity maxima (Girdler et al., 1969).

Rift zones of crevice type

These are characterised by grabens of great depth, e.g., the Lake Albert graben (2–3 km), the Upper Rhine graben (3–4 km) and the Baikal graben (5–6 km). The marginal uplift

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Fig. 1. The position of rifts and related zones in the tectonic structure of the earth.
Recent (late cenozoic) rift belts and zones. Continental rift belts: 1 = epiplatform arch-volcanic rift zones; 2 = epiplatform crevice-like rift zones; 3 = epiorogenic rift zones and belts; 4 = intercontinental rift zones. Oceanic rift belts: 5 = mid-oceanic ridges with axial rift valleys; 6 = mid-oceanic ridges without clearly expressed rift valleys; 7 = areas of mid-oceanic ridges with important volcanic manifestations; 8 = some large-scale faults and wrench faults active during the Cenozoic.
Pre-late Cenozoic zones of extension, fracturing of the crust and graben building – probable ancient analogues of recent rift zones (continental and intercontinental):
9 = Late Mesozoic and Early Cenozoic; 10 = Early Mesozoic; 11 = Paleozoic; 12 = Late Proterozoic; 13 = areas of Late Cenozoic volcanism (outside of alpine geosyncline orogenic belts); 14 = zones of Cenozoic epiplatform orogenesis (mountain building); 15 = recent deep sea troughs; 16 = recent geosynclinal zones; 17 = zones of Alpine and Laramide folding; 18 = zones of Mesozoic folding (Kimmerides, Nevadides etc.); 19 = zones of Paleozoic folding (Hercynides, Caledonides); 20 = zones of late-Proterozoic folding or regeneration (Grenvillides, Bailalides etc); 21 = Pre-late Proterozoic platforms; 22 = areas of the oceanic floor with sub-continental crust; 23 = deep sea depressions with sub-oceanic crust; 24 = oceanic basins with crust of oceanic type.

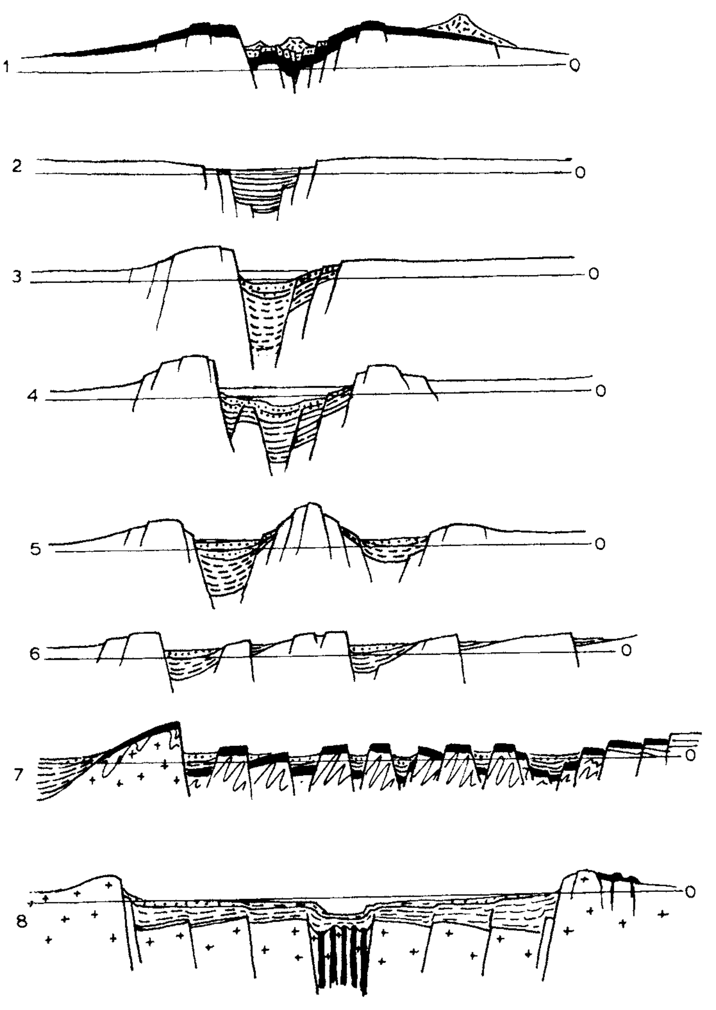

Fig. 2. The characteristic structural types of rift zones (the vertical scale of the cross-sections is exaggerated).
1 = arch-volcanic epiplatform rift zone (Kenyan rift zone according to B.H. Baker); 2–5 = crevice-like epiplatform rift zones: 2 = without marginal uplifts; 3 = with one marginal uplift; 4 = with two marginal uplifts; 5 = with internal uplift; 6 = zone of one-side tilted blocks; 7 = rift-like epiorogenic belt; 8 = intercontinental rift zone (rift zone of Red Sea according to C.L. Drake and R.W. Girdler).

zones are narrower and sometimes absent. Most are normal faulted and associated with horizontal extension. Some are associated with horizontal displacements and sometimes this component substantially exceeds the extension (e.g., the Levant Rift) although Razvaliaev (1971) suggests vertical movements are dominant. The rifts are seismically active and have narrow negative Bouguer anomalies due to to the thickness of unconsolidated sediments.

In contrast to arch-volcanic rifts, the crevice rifts have little volcanism (Tanganyika, Upper Rhine) or no volcanism (Baikal). The local volcanic centres are on the flanks and absent in the axial part of the graben. The volcanism takes place after graben formation and the concept of arch collapse cannot be applied to crevice-type rifts. The horizontal extension does not exceed 5–10 km although there is evidence for several kilometres thinning of the crust beneath the Rhine and Baikal Rifts. It seems that heat leakage is concentrated along narrow fissure zones rather than in the formation of large magma chambers as for the arch volcanic type rifts.

Rift zones of young folded regions

In contrast to the platform rifts, these occur in regions of folding following completion of the geosynclinal cycle. Examples include the Basin-and-Range rifts of the western U.S.A. (which followed a Mesozoic geosynclinal cycle) and the Early Mesozoic grabens of the west Siberian plate (which followed a Hercynian geosynclinal cycle). There is not one graben but a whole series of parallel grabens separated by narrow horsts, the relief having amplitudes of 2 to 5 km. The horizontal extension can be greater than 1000 km and there may be zones of horizontal displacement such as the San Andreas.

The formation of these zones was preceded and accompanied by intense eruptions of calc-alkaline magmas both acid and basic. The zones overlie a crust-mantle mix (Cook, 1966) and in time, follow the more heated and plastic conditions of an orogenic era.

The various types of rift zones are illustrated in Fig. 2.

RIFTS AND RIFT LIKE BELTS

All the rift belts in the ocean floor and the majority of those on the continents join to form the world rift system which has a length of many thousands of kilometres. Where rift belts join, there are usually zones of transform faults. The termination of the rift belts are often fan-like (e.g., Tanzania). However, there are isolated rift belts such as the Baikal rift zone and the southwest branch of the African rift system (Tanganyika–Rukwa–Nyasa).

INTERRELATIONS BETWEEN GEOSYNCLINAL-OROGENIC BELTS AND RIFT BELTS

The Cenozoic rift belts and synchronous Alpine geosynclinal-orogenic belts are usually separated by great distances and complement each other, the former being zones of extension and the latter zones of compression.

The interrelations can be more complicated. For example, the west Siberian early Mesozoic rift belt rests on the Ural–Siberian Palaeozoic fold belt. In an analogous way, the northern Mid-Atlantic Rift became superimposed on the Caledonian Grampian fold belt inheriting its trend. Thus, with time, orogenic belts can be transformed into rift belts. Conversely, rift belts can be transformed into geosynclinal depressions, an example being the Greater Donetz (Hercynian) Basin which was previously a Devonian rift zone.

PRE-CENOZOIC RIFT ZONES AND BELTS

More and more old structures analogous to recent (Cenozoic) rifts are being discovered in the continents. Of late Mesozoic age are the Saint Lawrence rift zone, the Davis Strait, the Canadian Arctic archipelago and the Niger (West Africa) system of Cretaceous grabens. Of early Mesozoic age are the rift belts of western Siberia, Hindustan, Mozambique straits and the Argentine shelf of South America. Examples of Palaeozoic rift zones include the Oslo graben, the Spitsbergen and East Greenland grabens and the Gulf of Suez. Today, these grabens are isolated but they may represent relics of ancient rift belts. It is remarkable that the majority are associated with the periphery of continents and break off at the margins of the Indian, Atlantic and Arctic Ocean basins. It is natural to assume that these represent offshoots of large rift belts which existed in the oceans during the Mesozoic and possibly earlier. The oceanic parts of the old rift zones have been lost in the process of ocean basin expansion.

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EAST AFRICAN RIFT DEVELOPMENT

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ABSTRACT

Logatchev, N.A., Beloussov, V.V. and Milanovsky, E.E., 1972. East African rift development. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15: 71–81.

Numerous age determinations on volcanic rocks from the Gregory Rift make it possible to represent its structural and magmatic evolution in six chronological stages (in m.y.B.P.): (1) 23–16; (2) 13.5–12; (3) 10–5; (4) 5–2; (5) 2–0.7; (6) 0.7–0. The initial rift trough appeared during the transition from the third to the fourth stage. Later it was substantially deepened and in some places narrowed and marginal faultsteps were formed.

In the development of the Western Rift, which is still poorly studied, two major stages may be distinguished: (1) an early stage of subsidence; and (2) a late stage of graben formation. With the latter is associated the accumulation of Villafranchian sediments of the Kaiso Formation.

Hypotheses which attempt to explain the origin of the East African rifts by the breaking-up and separation of continents tend to ignore the geological history of the development of the rift valleys. Even in regions which have undergone dicyclic rifting (Rungwe area, Shire, and Chilwa grabens) there is no evidence of any separation of the continent required by the hypothesis of sea-floor spreading.

INTRODUCTION

The present paper reviews some features of the formation of the East African rifts. It is based mainly on the results obtained by exploration carried out at the southern extremity of the Gregory (Eastern) Rift and in some parts of the Western Rift by the Soviet East African expedition in 1967–1969 with the friendly support of the senior staff of the Geological Surveys and Universities of Kenya, Tanzania, Uganda, Rwanda and Burundi.

During the last few years varying attention has been paid to the two rifts. The Gregory Rift, this bright structural "star", seems to have received more attention than the less striking Western Rift. The reason is that the peculiarities of structure and location of the Gregory Rift allowed investigators to expect the settling the most important problems of rifting within continents, as well as the ascertaining of the possible connection between continental and oceanic rifting. Historical reasons were also in favour of the Eastern Rift: complete geological mapping of its central and southern parts, as well as the results of stratigraphical and paleontological studies served as a good basis for further investigations. This

matter is well illustrated by the recent appearance of a great number of papers dealing with Gregory Rift geology and geophysics (Girdler et al., 1969; McKenzie et al., 1970; Saggerson, 1970; Searle, 1970; Williams, 1970; Wright, 1970; Baker et al., 1971; Baker and Wohlenberg, 1971, and others).

Nevertheless, the data concerning the Western Rift development are also of great importance and should not be ignored when discussing such a complex problem as continental rifting.

THE GREGORY RIFT

At present there are so many K/Ar dates on volcanic rocks of the Gregory Rift that it is necessary to systematize them and to accompany such a catalogue with a map showing the location of the dates*. The map (Fig. 1) contains more than 280 dates for 186 different localities in Kenya, Tanzania, and Uganda. It characterizes all the main stratigraphic units and includes some dates previously published by other writers (especially Bishop et al., 1969, and Baker et al., 1971) and over 50 new dates recently obtained in the U.S.S.R.

On the basis of K/Ar data as well as on geological grounds the process of rifting can be divided into six time-stratigraphic stages (in m.y.B.P.): (1) 23–16; (2) 13.5–12; (3) 10–5; (4) 5–2; (5) 2–0.7; (6) 0.7–0. Each stage reflects definite conditions of magmatism and structural development and is illustrated by schematic diagrams (Fig. 2) containing the most important information about the rifting process, such as the location of the major central volcanoes (the total number of which is over 70), fissure eruption belts, petrochemical associations of volcanics, and major faults along which visible vertical displacements took place by the beginning or during each stage. So the combination six stages considered together give the complex picture of migration in time and space of volcanicity and faulting which are the main geological expressions of rifting.

It should be noted that the chronological limits accepted are to a certain degree tentative and conventional. This is especially true of the limits of stage III and the lower limit of stage IV. The accumulation of new data will allow refinements to be made but is unlikely to change them in essence.

Stage I (23–16 m.y.B.P., Lower – Middle Miocene)

Stage 1 started with the subsidence of the Turkana depression and uplift of the area along the Uganda – Kenya frontier. It was accompanied by massive fissure eruptions of basalts, mugearites, and rhyolites in Turkana**. On the uplifted area the two neighbouring provinces of alkaline – carbonatite volcanoes of Eastern Uganda and the Kavirondo Rift were formed. In spite of their geographical proximity both provinces developed auton-

* The catalogue (in Russian) is in the possession of Dr. N.A. Logatchev, Institute of the Earth's Crust, Irkutsk 33 U.S.S.R.

** The age of Turkana rhyolites and mugearites is still uncertain. Some authors (Williams, 1970; Baker et al., 1971) believe that these rocks are of Early Pliocene age. Just as probably they may be thought to be late differentiates of Miocene volcanic rocks.

omously. The linear arrangement of the Miocene volcanoes Kisingiri, North and South Ruri, Homa, Buru (?), and pre-Tinderet coincides with the future strike of the Kavirondo Rift. In the south of Eastern Uganda Province central volcanoes were also grouped into a linear zone (Sukulu, Tororo, Bukusu, Sekululu) but north of Elgon they diverge into a broad zone of centres approaching the area of massive fissure eruptions in Turkana. Because of the absence of satisfactory stratigraphic dates there is some uncertainty whether flood basalts of residual plateaus to the north of Mt. Kenya (e.g., Merti plateau, Sagererua plateau) originated in Miocene or, as Williams (1970) believes, later in time. On account of some indirect evidence these basalts are here referred to the first stage (Early – Middle Miocene).

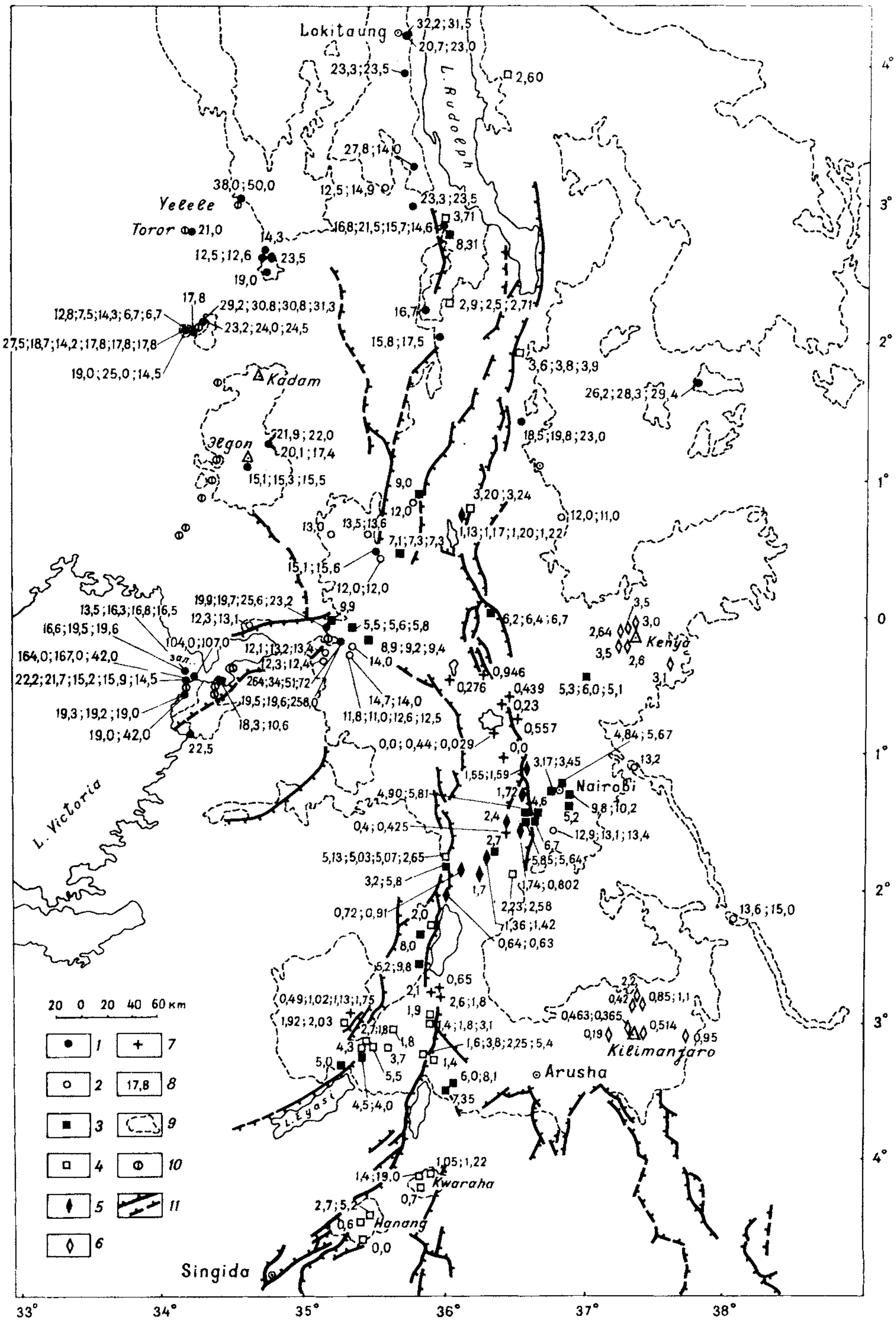
The discovery of continental Cretaceous sediments below Miocene basalts on the western coast of Lake Rudolf (Arambourg and Wolff, 1969) will possibly lead to the supposition that rift-forming movements began in the Cretaceous, but as the limits of distribution of dinosaur-bearing sediments and the formational style are uncertain, any opinion on this account is still open to question. In any case it would be premature to believe that sediments described as Turkana grits (Walsh and Dodson, 1969, and others) are all of Cretaceous age.

Stage II (13.5–12.0 m.y.B.P., Upper Miocene)

Stage II is especially important because it started events which pre-determined the location and development of the Gregory Rift structure. The area of volcanic activity shifted to the Kenya domal uplift and lost its connection with the Ethiopian volcanic area. Immense fissure eruptions of phonolites and phonolitic trachytes occurred through systems of fractures on a crest of an arched volcanic uplift along the strike of the future Gregory Rift and through additional branches of fissures. A vast, more or less symmetrical phonolitic shield was formed with a thickening along the culmination line in excess of 1000 m, gradually diminishing to the edges of the plateau. The main features of this stage are the dominance of fissure eruptions and the highest magmatic productivity during the whole history of rift development. The total volume of Upper Miocene phonolites is about 25,000–30,000 km^3. If single, and therefore possibly unreliable, K/Ar determinations are neglected, the majority of dates obtained in different laboratories of the world range from 13.5–12.0 m.y.B.P. Radio-isotopic stability and the monotonous petrochemical composition of plateau phonolites are evidence of an easy connection between the interior and the surface of the earth. Thus one can explain the high volcanic productivity of approximately 16,000–20,000 m^3/year.

Stage III (10–5 m.y.B.P., Lower – Middle Pliocene)

Stage III is the most complex, being characterized by radical changes in the distribution and character of volcanic activity although it appears at first to be a continuation and a


Lokitaung
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L. Rudolph
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Yeelee
23,3;23,5
Toror
21,0
14,3
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19,0;25,0;14,5
Kadam
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21,9;22,0
Elgon
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8,9;9,2;9,4
3,5
Kenya
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2,6
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0,439
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0,23
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18,3;10,6
11,8;11,0;12,6;12,5
0,557
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4,84;5,67
22,5
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0,0
13,2
1,55;1,59
3,17;3,45
L. Victoria
4,90;5,81
1,72
4,6
Nairobi
9,8;10,2
5,2
2,4
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12,9;13,1;13,4
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6,7
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1,7
0,72;0,91
2,23;2,58
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2,1
2,2
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2,6;1,8
0,42
0,85;1,1
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1,92;2,03
1,9
0,463;0,365
0,514
2,7;1,8
1,4;1,8;3,1
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0,95
4,3
1,8
1,6;3,8;2,25;5,4
Kilimanjaro
3,7
1,4
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5,5
Arusha
6,0;8,1
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L. Eyasi
1,05;1,22
1,4;19,0
Kwaraha
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2,7;5,2
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Hanang
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Singida
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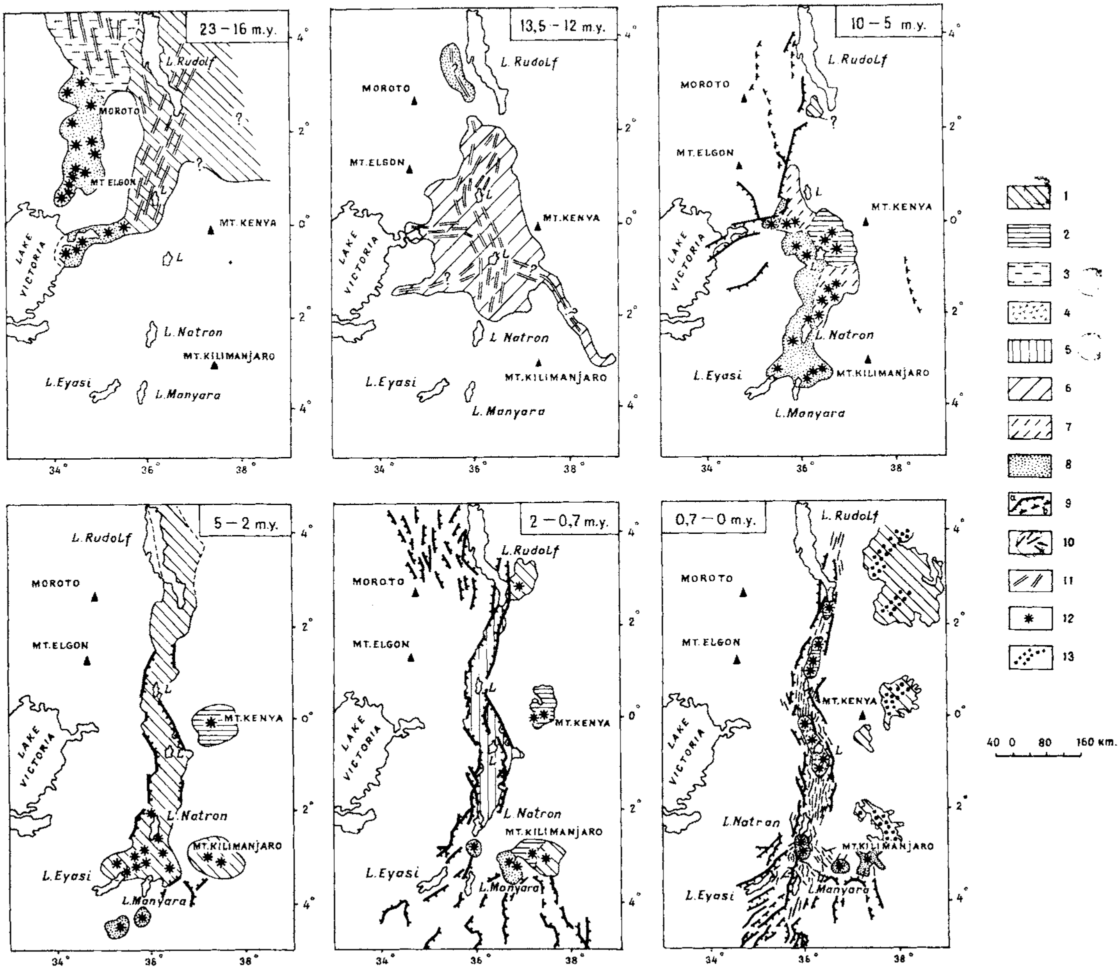

Fig. 2 (above). Miocene to Recent volcanicity and faulting of the Gregory Rift area. *1–8* = volcanic associations: *1* = basalt and basalt-trachyte; *2* = basalt–trachyte–phonolite; *3* = basalt–mugearite–rhyolite; *4* = trachyte–phonolite–comendite–pantellerite; *5* = trachyte–phonolitic trachyte–phonolite; *6* = phonolite; *7* = nephelinite – phonolite and basalt – trachyte; *8* = basanite – nephelinite – phonolite.
9 = faults: (a, real; b, inferred); *10* = later fissures on the rift floor; *11* = probable feeders of fissure eruptions; *12* = major central volcanoes; *13* = main fields of small vents in areas of fissure and/or multicentre eruptions (for stage 0.7–0 m.y. only).

Fig. 1 (opposite). Map of K/Ar dates (in m.y.) of lavas, tuffs and intrusive rocks. Dating groups: *1* = Lower and Middle Miocene; *2* = Upper Miocene; *3* = Lower and Middle Pliocene; *4* = Upper Pliocene (basalt series and its analogues); *5* = Upper Pliocene – Lower Pleistocene (trachyte series and its analogues); *6* = Upper Pliocene – Pleistocene (Kenya and Kilimanjaro volcanoes); *7* = Pleistocene (rift volcanics); *8* = dates in m.y.; *9* = Neogene – Quaternary volcanics; *10* = subvolcanic massifs (usually with carbonatites); *11* = faults.

concluding phase to the events of the previous stage. Volcanicity took place in a narrower zone but extended southwards beyond the southern limit of plateau phonolites at the latitude of Lake Magadi. A lack of dates makes it difficult to outline the northern limit of the area of Lower – Middle Pliocene volcanicity. At the southern termination of Lake Rudolf there are possibly some rare outcrops of volcanic rocks dating from 10 to 5 m.y.B.P. (Patterson et al., 1970), but as the limits of their areal distribution are uncertain, they can only be roughly outlined on a schematic diagram.

An essential peculiarity of this stage is the prevalence of eruption of the central type. The subsidiary fissure eruptions are located close to and in the rift valley (e.g., Shirere and Kishalduga melanephelinites, Nairobi phonolites and (?) trachytes, Thomson's Falls phonolites, etc.). Volcanoes, the total number of which is not less than 19, also have a close connection with the rift as they are situated along its floor (01 Esayeti, Olorgesailie, Lenderut, Mosonik) or over the scarps of the rift shoulders (Narok, Ngong, Niandarawa, Sattima, inferred centres in Mau range, Kilombe (?) Londiani, Tinderet). Though a deep control from the future rift valley is clearly expressed in the distribution of volcanoes, the rift valley is still not defined as a real rift trough.

Simultaneously with the transformation of the type of eruption from fissure to central type, the petrochemical composition of volcanic rocks changed substantially and became more variable. Nevertheless, in a larger part of the area (especially along the western side of the rift and in the south) the volcanicity preserved its strongly alkaline character. The alkalinity of the eastern associations is noticeably lower than that of the western ones.

At the beginning of and during the third stage, fault displacements took place along the Elgeyo Escarpment and to the north of it within a flexure zone separating the Turkana depression from the East Uganda upwarped area. Marginal faults of the Kavirondo Rift were possibly also formed at this time. The remaining part of the rift area was still not affected by noticeable fault displacements. Major boundary faults were primarily formed at the end of stage III or shortly after it (see Fig. 2, 5–2 m.y.B.P. stage). On the western side of the rift valley vertical displacements embraced nearly the whole length of the structure from Kamasia Range in the north to the Crater Highlands in the south. Displacements on the eastern side occurred only in a limited section at the foot of the Aberdare Range and west of Kajiado. So, a newly formed trough was built as an asymmetrical structure.

Stage IV (5.0–2.0 m.y.B.P., Upper Pliocene – Lower Pleistocene)

After the first rifting movements, massive basalt eruptions began at the bottom of the trough, as well as local accumulation of sediments. In the major part of the rift, basalts and associated minor trachytes were covered by later volcanics and sediments but undoubtedly they formed a continuous horizon spreading almost throughout the whole rift valley. The basalt series was formed mainly by fissure eruptions on the rift valley floor (Kirikiti basalts, Singaraini basalts, Kwaibus basalts), but at its southern extremity, where the latter expands like a bell, the eruptions were related to a group of shield volcanoes (Sambu, Gelai, Kitumbeine, Tarosero, large centres of the Crater Highlands).These centres partly remained ac-

tive during the next stage when the plateau trachyte series of the Magadi area and its analogues to the north were formed. Such a change of eruption type was probably conditioned by a dispersion of extension at the end of the rift valley just near the eastern edge of the archaic Tanganyika Shield. About 130 km southwards from the main basaltic area there existed a minor area of alkaline – carbonatite magmatism in which the formation of two phonolite – nephelinite volcanoes (Hanang and Kwaraha) was accompanied by numerous explosion craters. Thus, Upper Pliocene volcanicity was characterized not only by a change of eruption type along the rifting area, but also by some changes in petrochemical composition of magmatic products. In both respects the situation resembles the palaeovolcanic conditions of the Lower – Middle Miocene. These similar events differ only in areal expansion and geographical position.

Simultaneously with immense basalt volcanicity autonomous melting centres appeared to the east of the rift. They started the formation of the gigantic volcanoes of Kenya, Kilimanjaro and, probably, Kulal. The stage under review therefore appeared to be a turning point in the process of Neogene – Quaternary magmatism. The volcanic activity to the west of the rift stopped completely just at the start of rift down-faulting. Since that time volcanic activity was localized inside and to the east of the rift valley and the asymmetrical character of the deep magma generating process was revealed clearly for the first time.

Stage V (2.0–0.7 m.y.B.P.)

Stage V was a continuation of the previous one. Though spreading of volcanicity decreased, fissure eruptions, which formed plateau-trachyte series, still dominated on the rift floor. This is obvious at least for the southern sector of the rift from Gelai Volcano to Suswa Caldera. For a distance of 120 km the rift floor is covered by alkaline and quartz trachytes. North of Suswa the volcanicity was characterized not only by lavas of intermediate composition (e.g., Lake Hannington phonolites), but also by large masses of pyroclastics forming thick beds of ignimbrites and tuffs on the rift floor and on its shoulders, especially on the Mau Range (their coverage is not shown on the diagram on account of the high degree of material allochthony). North of Lake Baringo there existed probably some central volcanoes that erupted trachytes. Chepchuk Volcano is an example, as Carney et al. (1971) have shown.

The most important tectonic event of Stage V is rift-trough formation along its total length and intensive fracturing of the crust at its southern and northern extremities where systems of diverging faults formed with dominating eastward dip-slip displacement of the blocks. As this event took place during and subsequent to massive eruptions, it is believed that this beginning of a new phase of major rift floor down-faulting partly arose as a result of devastation of the interior below the rift. During this new stage of rifting, boundary faults of an earlier generation were renewed and lengthened along the strike, but the main event was the development of new faults inside the trough and at its extremities. Dislocations along them promoted deepening of the rift valley and the formation of step-faults at its margins. Thus, locally "rift in rift" structures with marginal platforms appeared.

Stage VI (0.7--0 m.y.B.P.)

When fissure eruptions stopped, the widespread area of volcanicity on the rift floor disintegrated into several separate areas of central volcanoes. The largest of them was connected with the middle segment of the rift between Suswa and Menengai calderas. Central eruptions of Kenya, Kilimanjaro, and Kulal nearly ceased and fissure and multi-centre outpourings, forming large basalt ridges and the basaltic Marsabit shield, appeared to the east of large cones. The development of these satellites caused partial or total eruptive extinctions of the main centres. Eastward migration of magmatic activity, having begun in the stage 5.0--2.0 m.y.B.P. retained its role during the last geological period. It is well illustrated by the volcano-pairs of Kilimanjaro – Chyulu, Kenya – Nyambeni, and Kulal – Hurri.

Simultaneously with the regional disintegration and transformation of fissure eruptions into central ones, the chemistry of volcanics of the rift floor diversified. Composition and location of volcanic associations testify to maximum heating in the central section of the rift valley. Melting in eastern chambers was deeper and caused formation of slightly differentiated associations, mainly of basaltic composition. It is essential to emphasize that strongly alkaline–carbonatite magmatism (Oldonyo Lengai, Kerimasi volcanoes) again demonstrated its connection with the southern extremity of the rift valley and the whole volcanic area. The extreme position of alkaline magmatic chambers reflects a general regularity that caused the creation of optimal thermodynamic conditions for the melting of deep horizons in the upper mantle at the end of the rift where mild extension favourably combined with thermal conditions.

Volcanicity and faulting in the Gregory Rift area show that in spite of its deep connections with the Ethiopian Rift and through it with the world-wide rift system, rifting in Kenya and Tanzania proceeded autonomously. The latter was defined in Late Miocene during massive eruptions, when the Gregory Rift volcanic area had lost its connection with the volcanic area of the Ethiopian dome. It was re-established only in Upper Pliocene at the beginning of the formation of the rift valley.

In terms of recent geological and geophysical data (McCall, 1967; Girdler et al., 1969; Williams, 1970; Baker and Wohlenberg, 1971) the central part of the Gregory Rift – the part of the valley which deviates westwards from the general strike of the rift – must be presumed to be its historical centre. This middle segment is peculiar in many respects: (1) it is the geometric centre of the rift valley and of the Late Miocene phonolitic cover; (2) the topographic culmination of the rift floor and shoulders coincides with it, and the surface of the Precambrian basement is deeply situated and probably represents a mirror image of the topography; (3) it is a zone of the highest volcanic productivity and permeability during the whole history of the rift development, which leads one to suggest a maximum replacement of the crust by deep injections and the highest position of the disturbed and thermally activated upper mantle with respect to the surface. This fact is confirmed by a positive peak on the long-wavelength negative gravity anomaly (McCall, 1967; Girdler et al. 1969). The probable cause of the specificity of the development of the central segment is a participation in the kinematics of left lateral movement along the strike of the rift valley.

It might be favourable to the creation of a zone of high permeability for igneous injections, heating of the substratum, and partial melting at lesser depth than at the extremities of the rift.

THE WESTERN RIFT

The ratio of volcanic to sedimentary rocks of the Western Rift differs from that of the Eastern one. Another peculiarity is that its southern part (Rukwa – Nyasa – Urema zone) first formed in Cretaceous times. After this its development stopped but started again in Neogene – Quaternary time, when rifting involved both the southern zone and the Tanganyika – Kivu – Semliki – Albert zone.

In the development of the Western Rift two stages may be distinguished: (1) early stage or a stage of preliminary subsidence during the Miocene and probably the Lower Pliocene; (2) late stage of the formation of rift grabens, commonly limited by boundary faults along both sides. The deposition of Kaiso sediments of Villafranchian age, which is still the most reliable single stratigraphic marker, refers to the beginning of the second stage. This two-staged history establishes a general similarity of the course and character of both branches of East African rift development, in spite of apparent differences in their morphology, Neogene – Quaternary fill, age, and tectonic type of their basement. The presence or absence of its connection with the world rift system does not obscure this general regularity

Volcanic rocks form some isolated areas along the Western rift and are associated with rises of the rift-trench floor. The Virunga area is active now and has been forming for 2–3 m.y. Toro–Ankole explosive eruptions began in Upper Pleistocene and stopped about 4,000 years ago. Southern Kivu and Kamituga basalt – trachyte volcanicity stopped in the Middle Pleistocene, but began at least 6–8 m.y.B.P. The Rungwe area volcanicity probably commenced at the end of the Miocene or at the beginning of the Pliocene, but its last eruption was at the very end of the 18th century (Harkin, 1960). Thus, the eruptions of all volcanic areas along the course of the Western Rift are arranged in a row, each northern member of which is younger and a shorter duration than the southern one. Such interrupted migration of surficial manifestations of magmatic activity along the Western Rift actually reminds one of a movement that is reverse to that of the migration of magmatism in the Gregory Rift area, which occurred successively in a N-S direction, from Ethiopia to Kenya and Northern Tanzania. In both cases the extremities (in the Western Rift in the north; in the Gregory Rift in the south) consist of very young or recent occurrences of strongly undersaturated lavas and carbonatites (Oldonyo Lengai and Kerimasi volcanoes of the Eastern Rift; fields of explosion craters and small cones of Toro-Ankole in the Western Rift).

CONCLUSION

The dynamic conditions of continental rifting are those of extension, which changes in time as well as along the rift zones. Estimates of displacements along boundary and rift interior faults, with planes inclined at an angle of 63–70°, hold only for the uppermost part

of the crust as many fissures inside the rift die out within a depth of a few kilometres. In any case, such shallow ruptures are "grid" faults, which produce an impression of considerable extension. Real extension of the whole crust must be less than that of its uppermost part where dip-slips are in part the result of domal upwarp. Moreover, the increase of the angle of dip of the fault planes with depth must be regarded as very likely.

The wide popularity of the hypotheses of sea floor spreading and plate tectonics stimulated some researchers, especially geophysicists (McKenzie et al., 1970; Searle, 1970), to apply them as an explanation of the Gregory Rift structural evolution. It is believed by the advocates of these hypotheses that the crust under the rift floor was disrupted and that its blocks were separated from each other by 20–30 km. But such a conclusion is based on insufficient use of available geological data, which compel one to believe that the Gregory Rift floor is composed mainly of Precambrian basement intruded by dykes, plugs, and plutons of basic and intermediate rocks. Along with high heating of the crust under the rift and saturation of the crust with material of deep-seated origin, partial melting and metamorphism of the original substratum toward denser rocks may have taken place. Such a complex of phenomena that made the crust denser despite extension and heating seems more acceptable than plate tectonics as an explanation of the gravimetric curve characteristic of the middle part of the Gregory Rift. This is undoubtedly a more complicated explanation than its alternative – disruption and separation of the crust by sea floor spreading – but it is based on real geological observations. The presence of Precambrian rocks in marginal fault-steps, xenoliths, and ejectimenta in lavas and pyroclastics of the rift valley, as well as the character of the basement exposed under the volcanic cover at both rift extremities indicate the presence of *a sialic basement under the rift.* Though the Gregory Rift has deep connections with the world rift system and though its formation was followed by strong volcanic action, it cannot be considered (as Baker and Wohlenberg correctly noted in 1971) an example of the first stage in the separation of a continent.

When discussing the mechanism of East African rifting it should also be pointed out that Precambrian basement rocks are exposed in some places of the Western Rift. In the Rungwe area, where the Rukwa, Nyasa, and Great Ruaha grabens join, we have the only opportunity to see the result of two-cyclic (Late Mesozoic and Neogene – Quaternary) rifting. The picture observed does not confirm the idea of crustal separation. The area between Nyasa, Shire, and Chilwa grabens, where the basement of a Late Mesozoic rifting complex is exposed, can be taken as another example. Large plutons of syeno-granites and nepheline-syenites as well as dyke swarms of alkaline rocks and carbonatitic bodies (Dixey, 1956; Garson, 1966) show that there was an area of high magmatic activity, the volcanic rocks of which were removed by erosion. But even this area does not provide any real evidence of crustal separation.

We must conclude that the hypothesis of disruption and separation of the crust, which is often used to explain the tectonics of ocean bottoms, cannot be applied easily to continental rifting. Any attempt to understand the origin of rifts on continents ought to take into account the real geological history. Geophysical information is of great importance for solving the problems of rifting, but it needs to be supported and controlled by geological data.

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THE KENYA RIFT VOLCANICS: A NOTE ON VOLUMES AND CHEMICAL COMPOSITION

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ABSTRACT

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Cainozoic alkaline volcanics associated with the Kenya Rift Valley have an estimated total volume of about 144,000 km^3. Basalts were erupted repeatedly and account for more than half this volume. The most strongly alkaline basalts occur in Miocene flows which flooded a proto-rift depression in northern Kenya. Post-Miocene basalts of the rift floor are less undersaturated (Niggli qz. value = –28) than those east of the rift (qz = –37). Some basalts at the rift margins show affinities with lavas of the rift floor; others more closely resemble rocks of multi-centre fields east of the rift. Nephelinites are frequently accompanied by highly undersaturated phonolites (phonolitic nephelinites) at central volcanoes in western Kenya (Miocene) and southern Kenya (Pliocene and Quaternary). Voluminous late Miocene phonolites (25,000 km^3) tend to be more undersaturated to the west than to the east. Excluding phonolitic nephelinites, post-Miocene phonolites of the rift floor are on average less undersaturated than phonolitic lavas outside the rift valley; they are also less undersaturated than the Miocene phonolites. Alkali trachytes, rhyolites and ignimbrites (10,000 km^3) which flooded the rift floor in southern Kenya and locally overflowed on to the plateaus were erupted during an important phase of domal uplift in Plio-Pleistocene times. Pliocene alkali rhyolites and ignimbrites in the northern part of the Kenya Rift evidently have a similar volume.

INTRODUCTION

The Kenya Rift Valley is part of the eastern rift system of Africa, and is characterized by widespread alkaline volcanism which took place more or less continuously from Lower or Middle Miocene times onwards.

Recent review articles have defined the main volcanic associations and have summarized some petrological aspects of the Kenya Rift volcanics (Williams, 1965, 1969a, 1970; King and Chapman, 1972). Others have described in very general terms the volcanic stratigraphy and geochronology (Bishop et al., 1969; King, 1970; Baker et al., 1971), or have outlined the structural evolution of the Kenya Rift (Baker and Wohlenberg, 1971). So far, petrochemical studies have been concerned with trends and variations only at individ-

ual centres or in parts of the volcanic field (Saggerson and Williams, 1964; Nash et al., 1969; Williams, 1969b; Macdonald et al., 1970; King and Chapman, 1972). The distribution of the volcanics and some aspects of their petrogenesis have been described in several accounts (e.g., Wright, 1963, 1965; Saggerson, 1970; Williams 1970) and attention has been drawn to the abundance of trachytic and phonolitic rocks; yet no estimates of volumes or proportions of volcanics of different compositions have been published to illustrate the magnitude of the petrogenetic problems involved.

The purpose of this contribution is to provide estimates of the volumes of volcanics in Kenya, and to draw attention to some general petrochemical characters and variations.

DISTRIBUTION AND VOLUMES

In considering the distribution and petrology of Cainozoic volcanics in Kenya, it is important to recognize essential differences between the widespread formations derived from fissure and multi-centre eruptions, and the lavas and pyroclastics of large central volcanoes (Williams, 1969a, 1970). The flood volcanics can be subdivided into a number of major stratigraphic groups, the ages of which have been more reliably determined than those of most central volcanoes (Baker et al., 1971). Moreover, the central volcanoes often display greater petrographic variety than do the more extensive flood volcanics, and they are frequently so complex that it becomes difficult to estimate the proportions of different rock types; this is a familiar problem, particularly in dealing with poorly dissected cones where the earliest flows may be completely unexposed. A problem of a similar kind is encountered in estimating volumes of volcanic infill within the rift valley itself. Some of the older formations, which are now exposed only in bounding fault scarps, were originally continuous and are now concealed in the rift floor. In the absence of deep boreholes and detailed geophysical studies, thicknesses of the older flows beneath the rift floor are largely conjectural.

The presumed distribution of volcanics in Miocene, Pliocene and Quaternary times is shown in Fig. 1, and the estimated volumes of flood volcanics are indicated in the explanation. Basalts include basanites, trachybasalts, mugearites and hawaiites as well as basaltic pyroclastics. Trachytic and rhyolitic tuffs, ashes and ignimbritic flows are grouped with lavas of these compositions. For simplicity, some slightly undersaturated Quaternary trachytes (phonolitic trachytes) are included with more common oversaturated varieties. The symbols in the explanation are not necessarily arranged in stratigraphical order.

Areas underlain by volcanics were measured systematically for each quarter-degree square, a distinction being made between products of major central volcanoes (subdivided broadly according to age) and those of fissure and multi-centre eruptions (subdivided on an age and composition basis). After making allowances for the effects of erosion on some of the older volcanoes, volumes of major centres were calculated by treating them as simple cones. Volumes of thinning sheets of lavas and pyroclastics derived from fissure and multi-centre eruptions were estimated by taking average thicknesses for the formations in

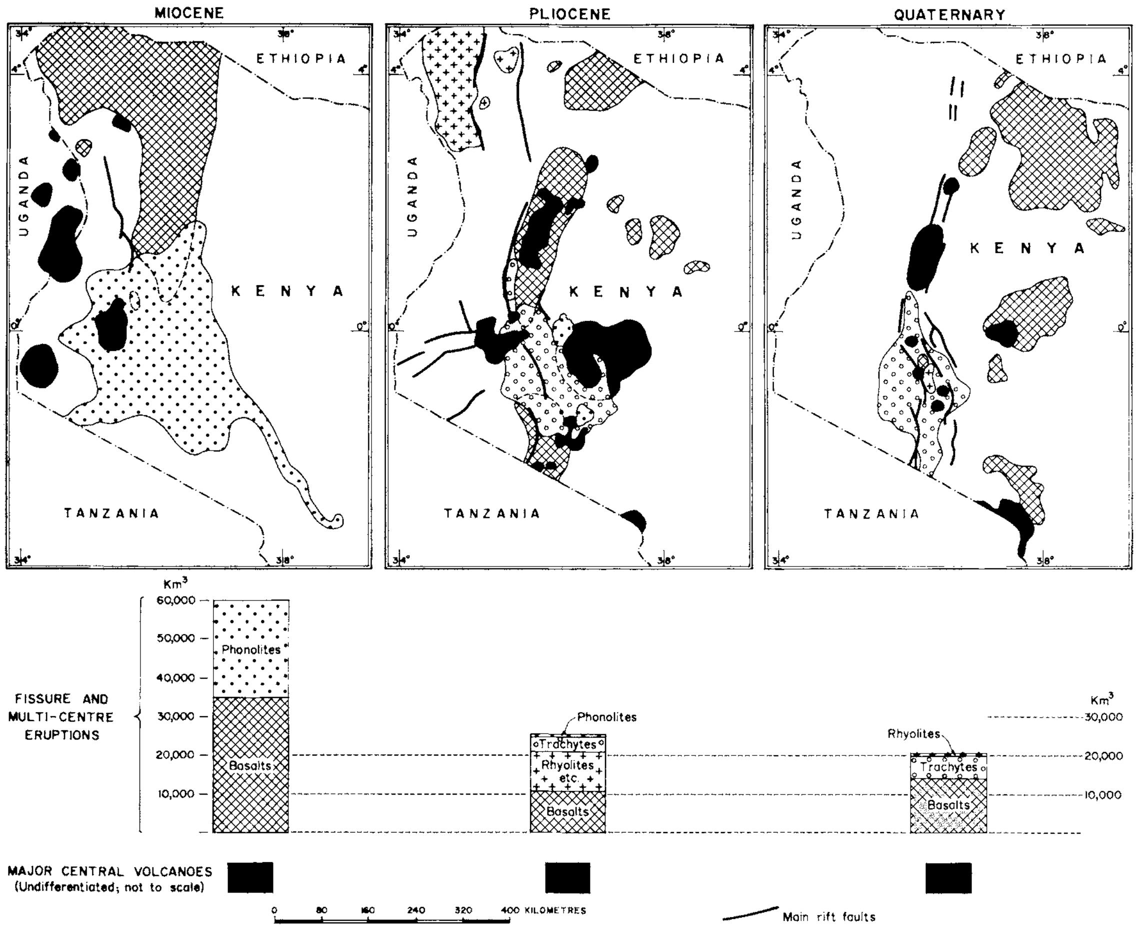

Fig. 1. Distribution and estimated volumes of flood volcanics and the location of major central volcanoes in Kenya.

each quarter-degree square. The thicknesses have been most reliably determined where the formations are exposed in major fault escarpments, or where the volcanics overlie well established erosion surfaces.

Miocene volcanics

The earliest basalts were erupted from numerous centres in a broad depression in northwestern Kenya, the development of which preceded the first phase of uplift in central Kenya in Late Miocene times (Baker and Wohlenberg, 1971). These volcanics have yielded isotopic dates in the range 14–23 m.y. (Baker et al., 1971) and they represent an extension of more prolonged and very widespread basaltic activity in Ethiopia (Mohr, 1968; Baker et al., 1972); there is no evidence to suggest that the flows ever extended south of the equator. With a total volume close to 35.000 km^3, the Kenya Miocene basalts belong to the most important single extrusive phase in the history of this part of the rift.

At about the same time as the eruption of the basalts in northern Kenya a string of large explosive central volcanoes developed in the Kenya–Uganda border area (King, 1965, 1970). These are composed mainly of nephelinitic lavas and pyroclastics, but at some centres alkaline intrusive complexes with carbonatites are exposed. The volcanics in Kenya probably exceed 5,000 km^3 in total volume.

In Late Miocene times (11–13.5 m.y.), before the main phases of rift faulting, updoming of south-central Kenya was accompanied by great eruptions of phonolites which flowed radially outwards from now concealed centres or fissures (Williams, 1970). These "plateau phonolites" have a total volume of about 25,000 km^3 and probably hold a unique position in records of alkaline volcanism.

Pliocene volcanics

Pliocene volcanism followed closely the line of a developing meridional trough bounded by faults and monoclinal flexures. The trough bisected the Kenya dome where major phases of uplift occurred during Late Pliocene and Early Pleistocene times.

Pliocene flood basalts, having a total volume of some 11,000 km^3, were more or less restricted to the rift floor where extensive eruptions evidently took place about 5 m.y. ago (Baker et al., 1971). Some of the lavas rest directly on Precambrian basement at rift margins near the Tanzanian border and must occupy much of the floor in southern Kenya. The Mid-Pliocene age of thick basalts within the rift in central Kenya is now well established (King and Chapman, 1972), but the distribution of Pliocene lavas towards the Ethiopian border has yet to be determined. East of the rift, thin basaltic flows of presumed Pliocene age (Williams, 1966) rest on sediments and Precambrian gneisses.

Rhyolites, ignimbrites and mugearitic lavas which overlie Miocene basalts in northwestern Kenya (Walsh and Dodson, 1969) are probably of Early Pliocene age, and may be related to much more extensice silicic volcanism reported in Ethiopia (Mohr, 1968). The

Kenya flows, which have been mapped only in rapid reconnaissance style, evidently have a total volume of 10.000 km^3.

Phonolitic flood volcanism occurred in two areas at the rift margins in central and southern Kenya (Fig. 1), but these Mid-Pliocene (5–7 m.y.) flows do not exceed a few hundred cubic kilometres in total volume.

During the Pliocene some 4.000 km^3 of trachytic volcanics were erupted from a down-faulted axial portion of the Kenya dome. Some trachytic lavas are of Mid-Pliocene age (7 m.y.), but more important are ignimbrites with intercalated lavas (2.5 – 5 m.y.) which flooded the rift floor and overflowed locally on to the marginal plateaus; they form the lower part of a Plio-Pleistocene trachytic group (Williams, 1970; Baker et al., 1971).

No attempt is made here to distinguish compositions of lavas and pyroclastics of major central volcanoes, for the main associations have been described elsewhere (Williams, 1969a, b; 1970) and attention drawn to the occurrence of two alkaline series (basalt–trachyte–phonolite and nephelinite–phonolite) at some of these centres. Pliocene central volcanoes probably contributed some 31,000 km^3 of volcanics of various compositions.

Quaternary volcanics

Voluminous trachytic volcanism, which commenced with lava and ignimbrite eruptions during the Pliocene, continued into the Pleistocene when some 6.000 km^3 of trachyte lavas and subordinate ignimbrites spread across the rift floor in southern Kenya. Later Pleistocene and Holocene activity was also predominantly trachytic and was associated with the development of a string of caldera volcanoes along the rift floor. Rhyolitic flows, having a total volume of only a few hundred cubic kilometres, occur in parts of the rift floor south of the equator; they are mostly of Late Quaternary age.

At about the same time as the eruption of trachytic and rhyolitic volcanics in the rift floor, activity in large multicentre fields east of the rift valley led to the accumulation there of 14,000 km^3 of basaltic lavas and pyroclastics.

Volcanics associated with Pleistocene and Holocene major centres have a total volume of some 3,000 km^3.

CHEMICAL COMPOSITIONS

A distinction is made between Miocene and post-Miocene volcanics in the alkalis–silica diagram (Fig. 2) in which 222 analyses are plotted. Approximate contents of silica and total alkalis in lavas having little or no nepheline are given in Table I, together with a subdivision, based on silica and total alkali content, of lavas containing more conspicuous nepheline.

The dashed line in Fig. 2 separates post-Miocene nephine-rich lavas from nepheline-poor and nepheline-free varieties. A less satisfactory line (dotted in Fig. 2) can be drawn for the Miocene volcanics, partly because of lack of data from nepheline-free intermediate

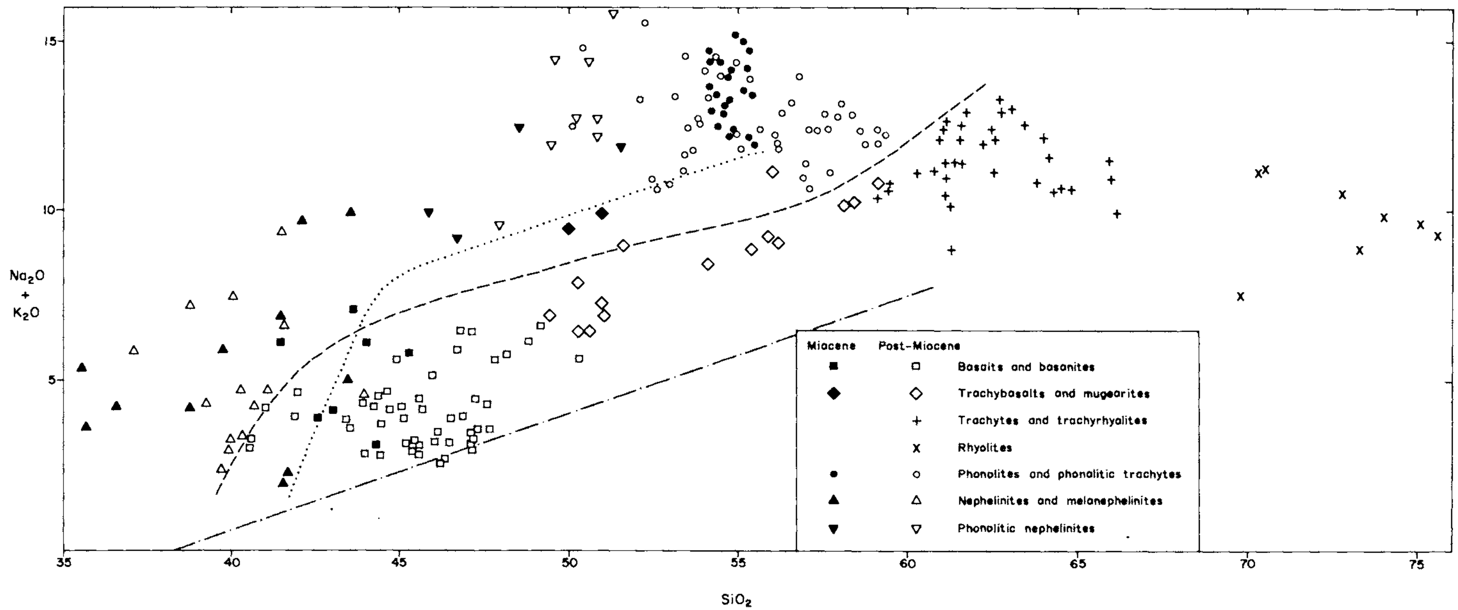

Fig. 2. Alkalis–silica diagram for the Kenya rift volcanics. The dashed line separates post-Miocene nepheline-rich lavas from those containing little or no nepheline. A similar division of the Miocene volcanics is made by the dotted line. The third line represents the boundary between the alkalic and tholeiitic suites in Hawaiian volcanics.

TABLE I

Subdivision of Miocene and post-Miocene lavas using silica and total alkalis content

	SiO_2 (%)	$Na_2O + K_2O$ (%)
Lavas with little or no nepheline		
Basalts and basanites	40 – 50	2.5 – 7
Trachybasalts, mugearites and hawaiites	50 – 59	6.0 – 11.5
Alkali trachytes and trachyrhyolites	59 – 66	8.5 – 13.5
Alkali rhyolites	69 – 76	7.5 – 11.5
Lavas with more conspicuous nepheline		
Nephelinites and melanephelinites	35 – 45	2 – 10
Phonolitic nephelinites	45 – 52	9 – 16
Phonolites and phonolitic trachytes	52 – 60	10.5 – 15.5

lavas. The thoroughly alkalic nature of the Kenya volcanics is illustrated by the fact that they all plot above the line representing the boundary between the Hawaiian alkalic and tholeiitic suites (Macdonald and Katsura, 1964).

Basaltic, trachytic and phonolitic volcanics are now examined in a little more detail to show some variations with age and location.

Basalts

Basaltic volcanics have been subdivided into four groups in the alkalis–silica diagram (Fig. 3). The pre-rift Miocene basalts tend to be more undersaturated and more strongly

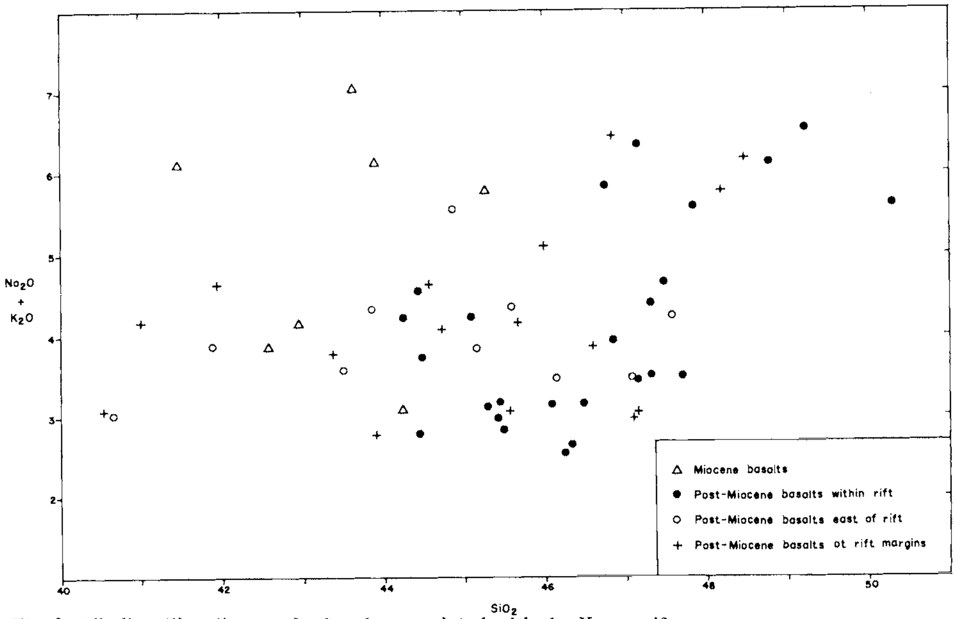

Fig. 3. Alkalis–silica diagram for basalts associated with the Kenya rift.

alkaline than most of the later basaltic flows. Post-Miocene basalts within the rift range in age from Pliocene to Holocene, but most analysed specimens come from Pliocene flows. They all contain more than 44% SiO_2 and are never as undersaturated as some of the Quaternary basalts east of the rift. The rift basalts have 2.5–6.5% total alkalis, but most of the basalts outside the rift fall in the narrower range 3–4.5%.

Post-Miocene basalts at the rift margins are associated chiefly with major central volcanoes. They exhibit a very wide range in chemical composition, sometimes even at a single centre. At the Aberdare volcano, for instance, basalts have SiO_2 contents ranging from less than 41% to more than 48%, but there is a suggestion that early lavas have affinities with those of the rift floor whereas later flows more closely resemble basalts erupted outside the rift.

Trachytic volcanics

All the analysed trachytic lavas plotted in Fig. 4 came from localities in the rift floor or from rift margins. Some are the products of fissure and multi-centre eruptions; others are clearly related to major centres, including caldera volcanoes.

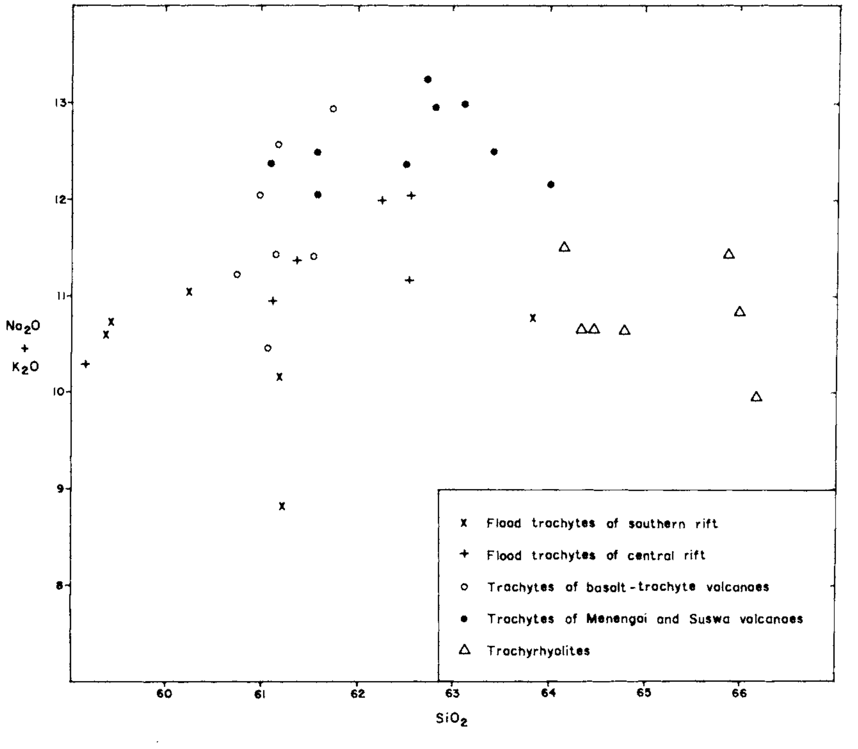

Fig. 4. Alkalis–silica diagram for trachytic volcanics of the Kenya rift.

Flood trachytes of the central part of the rift tend to be more alkaline than those of the southern section. Both groups grade locally into more silicic volcanics, some of which have been distinguished here as trachyrhyolites (chiefly pantelleritic trachytes). Macdonald et al. (1970) have drawn attention to the fact that alkalis are more likely to be accurately preserved in fresh glasses, and it is significant that the glassy trachytes of Menengai and Suswa caldera volcanoes have higher total alkalis than most of the other analysed specimens.

Phonolitic lavas

The SiO_2 content of phonolitic rocks ranges from less than 46% to over 59%, but an important distinction between nepheline-rich phonolites (phonolitic nephelinites) and most other varieties can be made at 52% SiO_2 (Fig. 5).

Miocene plateau phonolites fall in the narrow range 54–55.5% SiO_2. Total alkalis vary from about 12–15%, but the lavas west of the present rift valley (kericho, Mau, Mara, Kisumu and Uasin Gishu phonolites) tend to be more strongly alkaline than flows of the same age east of the rift (Rumuruti and Kapiti phonolites).

Post-Miocene phonolites of the rift floor contain about 56–59% SiO_2, whereas those west of the rift have less than 54% SiO_2. The analysed phonolites from localities east of the rift show a very wide range in SiO_2 content (50–59%); they come mainly from the large central volcanoes of Mt.Kenya and Kilimanjaro.

SUMMARY AND CONCLUSIONS

Estimated volumes of volcanics in Kenya are summarized in Table II. The total of 144,000 km^3 can be compared with a figure of 345,000 km^3 quoted for Ethiopia (Mohr, 1968) where the volume ratio of basaltic to silicic volcanics is approximately 6: 1. In contrast, the ratio of basalts to intermediate and silicic types among the flood volcanics in Kenya is about 1.3 : 1. Moreover, trachytic and phonolitic products at major central volcanoes appear to be at least as abundant overall as basalts and nephelinites. Calculations suggest that the volume ratio of basic volcanics to those of salic and silicic compositions has not been constant with time. Much more evidence is still required, but the following provisional figures show that the volume relationships cannot be ignored in discussions of petrogenesis:

	basic	:	salic/silicic
Miocene	1.5	:	1
Pliocene	0.7	:	1
Quaternary	2.1	:	1

The most voluminous eruptions of phonolites (25,000 km^3) took place towards the end of the Miocene; and alkali trachytes, rhyolites and ignimbrites having a total volume of the

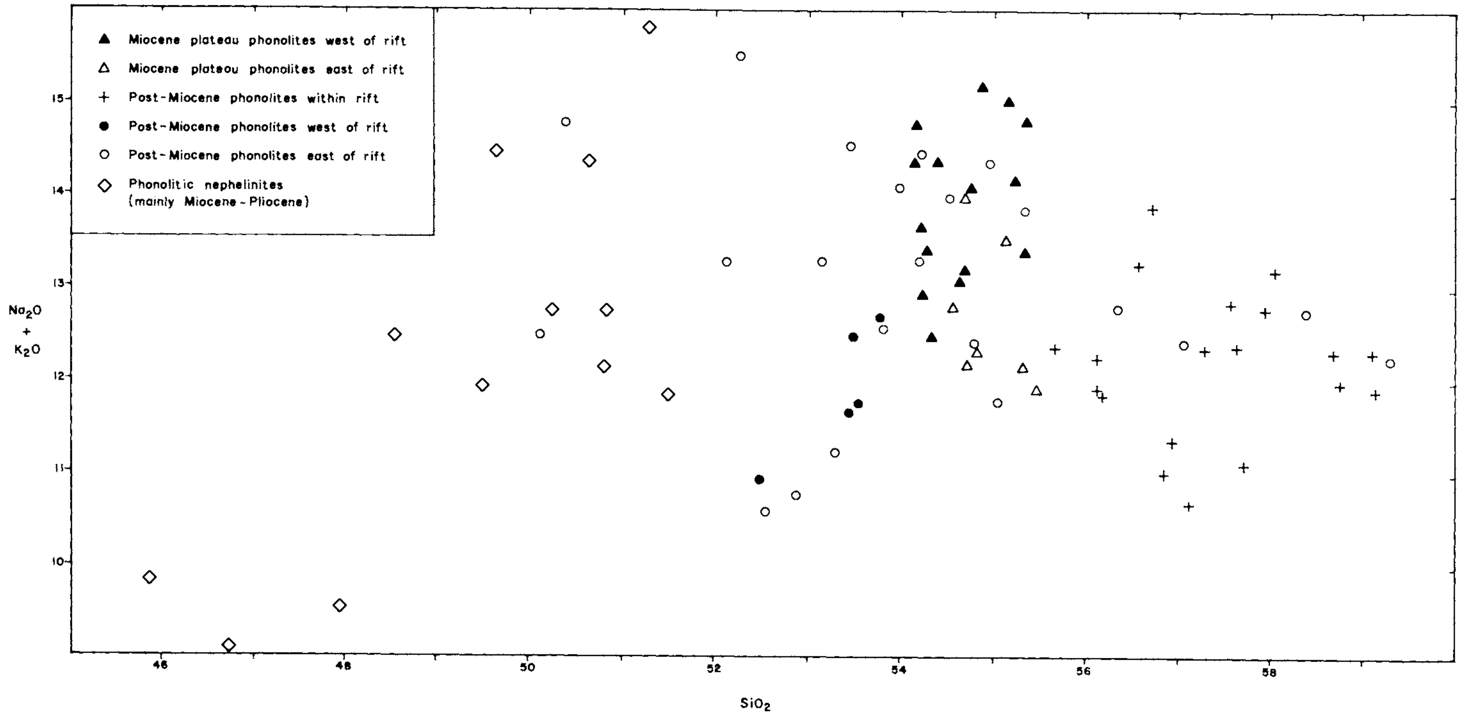

Fig. 5. Alkalis–silica diagram for phonolites, phonolitic trachytes and phonolitic nephelinites from Kenya.

TABLE II

Location of volcanic activity and approximate volumes and rates of accumulation of volcanics in Kenya

Age	Location of main volcanic activity		Volumes of flood volcanics (km^3)				Volumes of central volcanoes (km^3)	Total volume (km^3)	Approximate rates of accumulation (km^3 per m.y.)
	flood volcanism	major central volcanoes	basalts	trachytes, rhyolites, ignimbrites	phonolites	total			
Quaternary (0–2.5 m.y.)	within rift and east of rift	within rift and east of rift	14,000	6,000	minor phonolitic flows included with trachytes	20,000	3,000	23,000	9,200
Pliocene (2.5–12 m.y.)	within rift	within rift and east of rift	11,000	14,000	few hundred	25,000	31,000	56,000	5,900 (12,500) *
Miocene (12–23 m.y.)	proto-rift zone	west of rift	35,000	rare	25,000	60,000	5,000	65,000	5,900
			60,000	20,000	25,000	105,000	39,000	144,000	

* Rate of accumulation assuming all Pliocene activity occurred during the period 2.5 – 7 m.y.

TABLE III

Some petrochemical characters of basalts and phonolites based on average compositions

Age and location	Basalts					Phonolites (excluding phonolitic nephelinites)				
	number of analyses	average SiO_2	average alkalis	Na_2O/K_2O	Niggli quartz number	number of analyses	average SiO_2	average alkalis	Na_2O/K_2O	Niggli quartz number
Miocene	7	43.5	5.2	2.7	– 43	7	55.0	12.6	1.4	– 56 (eastern outcrops)
						15	54.8	13.9	1.4	– 72 (western outcrops)
Post-Miocene, rift floor	24	46.4	4.2	2.8	– 28	17	57.4	12.1	1.4	– 42
Post-Miocene, east of rift	10	44.6	4.0	3.0	– 37	20	54.0	12.7	1.6	– 64
Post-Miocene, rift margins	15	44.9	4.0	2.3	– 33	–	–	–	–	–
Post-Miocene, west of rift	–	–	–	–	–	5	53.5	11.9	1.2	– 53

order of 30,000 km^3 were erupted in central Kenya in Plio-Pleistocene times. These episodes of salic and silicic volcanism coincided with important phases of domal uplift (Baker et al., 1971).

The approximate rates of accumulation of volcanics quoted in Table II are based on the assumption that volcanism was more or less continuous. There is, however, some evidence to suggest that most of the Pliocene activity took place during the period 2.5 to 7 m.y.; in which case the Quaternary represents a time of waning volcanism.

Some petrochemical characters of basalts and phonolites are summarized in Table III, and the following points are worthy of note.

(1) Post-Miocene basalts and phonolites (excluding phonolitic nephelinites) of the rift floor are less undersaturated than flows of corresponding compositions erupted outside the rift. In the case of the basalts, this is probably related to shallower melting under the rift zone than beneath the plateau to the east. It should be noted, however, that the basalts of the rift floor do not show the strong tholeiitic tendencies found in rift basalts in Ethiopia (Mohr, 1971).

(2) Miocene basalts and phonolites tend to be more undersaturated and more alkaline than post-Miocene flows.

(3) Miocene phonolites exposed west of the rift are more undersaturated and have on average higher alkalis than flows of the same age on the eastern side of the rift; Na_2O/K_2O remains constant.

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A GEOCHEMICAL STUDY OF SILALI VOLCANO, KENYA, WITH SPECIAL REFERENCE TO THE ORIGIN OF THE INTERMEDIATE–ACID ERUPTIVES OF THE CENTRAL RIFT VALLEY

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ABSTRACT

McCall, G.J.H. and Hornung, G., 1972. A geochemical study of Silali volcano, Kenya, with special reference to the origin of the intermediate–acid eruptives of the Central Rift Valley. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 97–113.

The Rift Valley zone in East Africa may be regarded as an aborted ocean of long existence but never significantly opened up. Distensive regimes of tectonic up-doming, eruptivity and subsidence on normal faults, and median positive gravity anomalies complicating the wide negative gravity anomaly of Bullard must now be fitted into any model for the Rift Valley.

The petrological and geochemical nature of the eruptives is equally important: it is common to equate the abundance of salic volcanics in any section with thickness of the sialic crust there, assuming sialic remelting or sialic contamination.

The results of a detailed geochemical study of a single Quaternary caldera volcano, Silali, and its Plio-Pleistocene volcanic foundation are summarised here, and used as a model for differentiation in the central Gregory Rift Valley. A monistic differentiation theory must be sought to explain the peralkaline sialic eruptives, associated with alkalic basalts in great volumes, from the Miocene onwards. Relative to the basalts there is an increase in Si, Na, K and depletion of Mg, Ca, Ti; total iron shows slight depletion and the iron shows oxidation to the ferric oxide. Of the minor elements, Rb, Zn, Y, La, Zr, Pb, and Nb are increased; Cu, Ba and Sr depleted; and Cr, Co and Ni show slight depletion. The trends seen in trachytes, etc., are incipiently evident in hawaiites and mugearites. The consistent bimodality, peralkaline character of the intermediate to acid rocks, major and minor element contents, and intimate intermingling of the two suites in space and time, can only be interpreted in terms of complex differentiation process affecting a mantle-derived basalt parent.

Comparisons are drawn with the volcanic suites of Tenerife and with a differentiated intrusion in Australia which shows bimodality as a product of differentiation. It is suggested that a chain of cupolas underlie the Rift Valley, and in these differentiation follows an agpaitic regime, under the control of volatile concentration. Such cupola reservoirs probably reach up close to the floor of the Rift Valley. The change from phonolite dominance to trachyte–comendite dominance in the intermediate suites is the only significant time progressive change evident from Miocene to Recent. Notwithstanding this, the intermediate–acid rocks, and basic rocks associated with them plot consistently on FMA and alkali silica diagrams throughout this period.

INTRODUCTION

The current interest in plate tectonics and continental dispersion has resulted in much theorising concerning the nature of the Rift Valley in Kenya and Ethiopia. Evidence concerning the nature of the crust and upper mantle beneath the rift zone is critical to such theorising. It is now widely accepted that there can never have been any significant opening up of this particular rift: the concept of *an aborted ocean* (McCall, 1965) appears to be realistic, applied to the Great Rift Valley as a whole, from South Africa to the Red Sea. The once favoured compressional model of Bullard (1936), based on the recognition of a negative gravity anomaly and involving ramp faulting, can neither be reconciled with the observed tectonic regime of up-doming, eruptivity and subsidence on a multitude of normal faults, nor with the complication of a median positive Bouguer gravity anomaly in Kenya (McCall, 1967; Searle, 1970) and Ethiopia (Gouin and Mohr, 1964). The new gravity data, combined with the results of systematic regional mapping over two decades, have introduced a completely new factual basis for models of the development and geotectonic infrastructure of the rift. No less important than the geophysical data is the accumulation of petrological and geochemical data. It has been suggested that variations in the nature of the eruptive suites from one sector of the rift to another along its length might reflect variations in the nature of the underlying crust and mantle: in particular, abundant intermediate–acid volcanics might reflect a thick sialic crust, a paucity of such volcanics might reflect a thin sialic crust. This suggestion leans heavily on hypotheses of sialic remelting. It is the aim of this contribution to the symposium to enquire into the nature of the intermediate–acid rocks, particularly in relation to the validity of this supposition. The results of a very detailed geochemical study of Silali volcano situated to the north of Lake Baringo are summarised, and utilised as a basis for discussion of this question.

THE ERUPTIVITY OF THE CENTRAL GREGORY RIFT VALLEY

The eruptivity of the Gregory Rift Valley close to the Nakuru–Naivasha culmination – the meridionally elongated dome of maximum upwarping of the continental surface and highest standing sector of the Rift Valley floor – is quite distinct from the eruptivity that characterises areas peripheral to this culmination, where extensive nephelinite and melilite lava and pyroclast sequences occur. In the vicinity of the central culmination, alkalic basalts occur together with intermediate trachytes and phonolites, and subordinate acid comendites. The plutonic enclaves in the volcanics, and intrusive bodies laid bare in the underworks of volcanoes by dissection, are of ijolite, uncompahgrite and foyaite character in the outlying areas (McCall, 1957c), whereas they are of gabbroic and syenitic character in the central area (McCall, 1967, 1970). The outlying nephelinitic suites do appear to represent a quite separate genetic line of descent (Williams, 1971); V.I. Gerasimovsky (personal communication, 1971) confirms this, having delineated two quite discrete lines of descent on variation diagrams shown to Williams and the first author. The

eruptives of the central suites display a simple essential character, despite complications of a minor order, and appear to represent a single genetic suite, plotting on a single line of descent from an alkalic basalt parent, in contrast to the nephelinitic parent of the outlying suites.

Great volumes of intermediate to acid volcanics – trachytes, phonolites and comendites – are associated with smaller volumes of alkalic basalts. Despite the evidence of Williams and Gerasimovsky that this is a single line of descent from a basalt parent, the volume relationships are anomalous, and there is distinct bimodality. Something more than simple Bowen-type fractionation is needed to explain the observed relationships. The bimodality is evident both in the volcanic and plutonic rocks, for enclaves of gabbro in the basalts and of syenite in the trachytes correspond to the two contrasting volcanic composition ranges. So marked is the volume anomaly that immense caldera volcanoes, Menengai (McCall, 1957a, b, 1963, 1967) and Suswa (McCall and Bristow, 1965; Johnson, 1969) show no trace at all of basalt effusives, being entirely composed in their visible expression of intermediate rocks.

One may approach the problem of derivation of the intermediate and acid eruptives from the basalt stem in several ways. Each eruptive type may be considered in isolation and in detail – for example the pantellerites (MacDonald et al., 1970). In extreme contrast to that approach is that of Williams (1971) and Gerasimovsky, embracing the eruptives of the entire province. The authors have adopted a third approach, using the suite of eruptives in a restricted area as a test case for a study of relationships that apply to the entire range of eruptive suites of the Gregory Rift Valley near the Nakuru–Naivasha culmination. They believe that the results obtained from Silali and its sub-volcanic foundation of similar eruptives can be extended to the problem of the origin of the entire range of intermediate–acid eruptives associated with alkalic basalts in the Gregory Rift Valley.

Silali volcano, which has not previously been studied (McCall, 1968a, b, 1970), presents an alternating sequence of basalt and trachyte/comendite flows and tuff horizons. The two types of emission were intimately intermingled in time and space from Plio-Pleistocene to historic times. The rocks of the foundation beneath the volcano are very similar to those that make up the volcanic pile, and are probably equivalent to those of the Kinangop (McCall, 1967) – that is, about 2–3 m.y. old. Eruption in Silali itself commenced at some time equivalent to the Kariandusi sedimentation ("Kanjeran": McCall, 1967; McCall et al., 1967) – that is, about half a million years ago. These are only estimates, for there is a dearth of age dating results on rocks of this sector of the Rift Valley. The area is particularly suitable for a study of this kind on account of the wealth of "plutonic" enclaves in the volcanic rocks, representing hidden intrusives that have no surface expression.

Viewed in time we have, in this area, the upper part of a sequence of eruptions that extended from Miocene to historic times. The entire sequence displays bimodality and anomalous volume relationships. Though the earliest intermediate volcanics are phonolites, with subordinate trachytes, whereas the later intermediate volcanics are trachytes with subordinate comendites and phonolites, the basalts are similar, and the fundamental problem of the origin of the observed bimodality and volume anomaly concerns the entire

sequence. There is no possibility that a different genesis produced the basalt–phonolite bimodality to that which produced the basalt–trachyte/comendite bimodality considered here. There is only limited evidence of progressive development from Miocene to Recent times – there are more limburgites and picrite–basalts in the oldest basalt sequences; phonolites rather than trachytes or comendites predominate in them, and quartz–trachytes/comendites characterise the Plio-Pleistocene suites.

Viewed in space we see an intermingling of basic and intermediate/acid volcanics throughout this long history of eruptivity; the degree of intermingling is particularly marked in the later suites, such as that of Silali. There can be little doubt that reservoirs supplying feeders with magma of the two contrasting types remained active, discrete and available over most if not all of the period following the initial Miocene eruptivity (McCall, 1968b).

There is a mineralogical hiatus between the basic and intermediate/acid suites and a chemical hiatus, despite this intimate spatial and chronological intermingling: although there are minor developments of linking chemical types elsewhere in the Rift Valley province of Kenya, no such linking types are known here. Even where basic and intermediate/acid lavas have erupted alternately from a single minor volcanic centre, they maintain their completely distinct chemical and mineralogical character.

SUMMARY OF THE GEOLOGY OF SILALI

The foundation upon which Silali rests comprises older intermediate/acid volcanics – trachytes, comendites, rare phonolitic trachytes and phonolites. Subordinate olivine basalts come in quite high in this sequence. Tuff lavas (or froth-flows, McCall, 1964) are abundantly developed in the intermediate volcanic sequences, together with pumice tuffs. The tuff lavas range from eutaxites to confused autobreccias, containing trachyte blocks metres across. Syenite enclaves are common in these tuff lavas and in the pumice tuffs; the tuff lavas are regarded as ground-hugging, turbulent and laminated, fragmental effusions, not tephra.

The volcano of Silali covers more than 300 km^2 and the summit stands more than 700 m above the floor of the Rift Valley. It is a composite volcano, a dome built from clustered vents: the first phase of trachyte effusion was followed by emission of trachyte pyroclasts, which, in turn, was followed by a return to quiet effusion. Later, the volcano suffered sagging along a median, meridional zone, in which a fine grid of normal faults developed immediately after extensive emission of thin basalt flows from many small vents situated along these fracture lines. The faulting is the closest-spaced grid of McCall (1967) and McCall et al. (1967), probably Holocene, and part of Gregory's "Clapham Junction" pattern. The basalts erupted obscured the trachytes of the main phase, but further trachyte eruption then occurred just prior to the formation, by subsidence, of an oval, 8 × 5 km caldera: in the last events following the formation of the caldera basalt and trachyte have been erupted from minor, scattered centres, mainly situated outside the caldera.

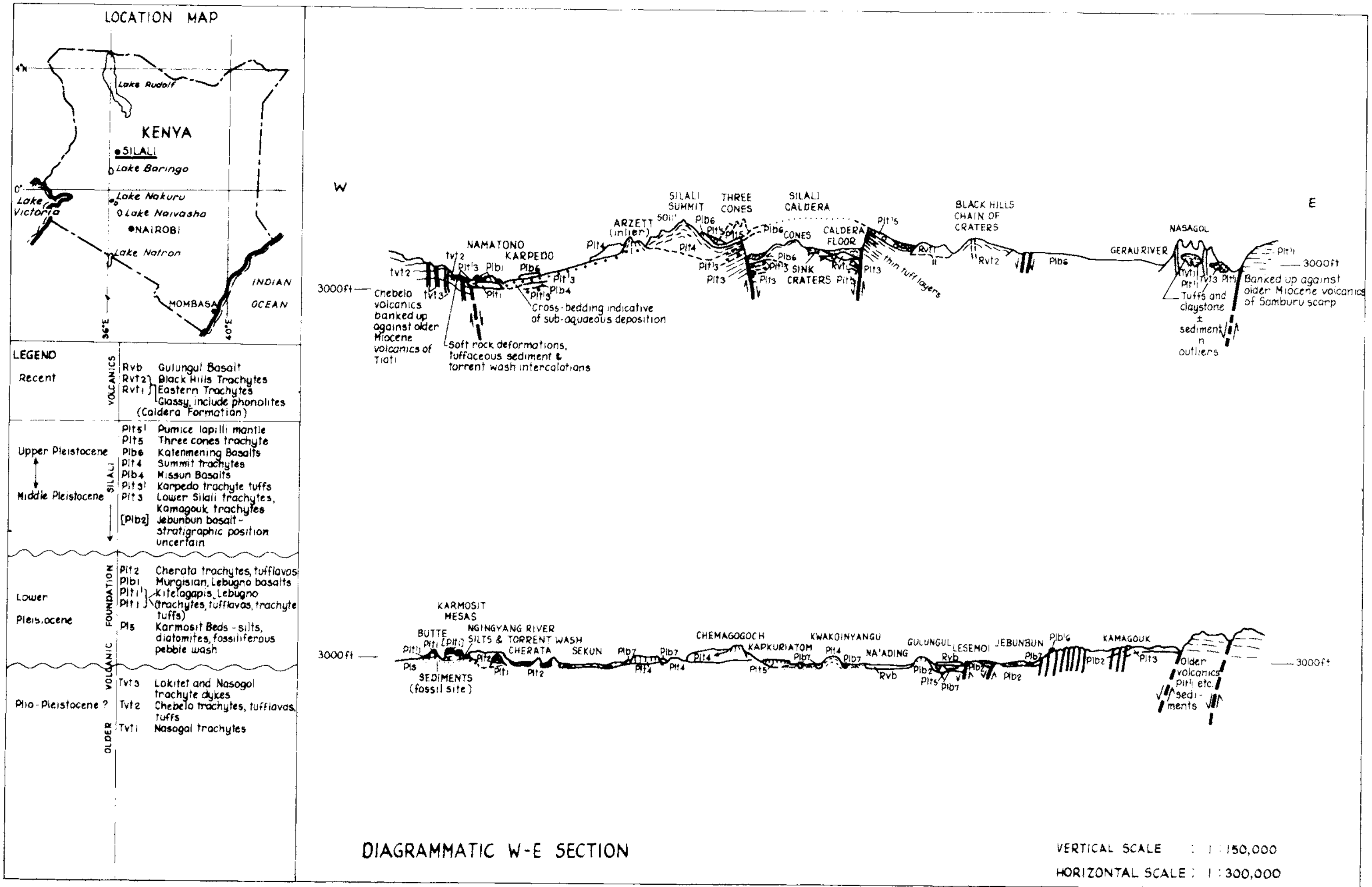

Fig.1. Location of Silali and sections across the area, showing the formations from which specimens were collected for analysis.

TABLE I

Major elements

	1	2	3	4	5	6	7	8	9
SiO_2	46.95	51.10	53.30	51.00	51.10	47.97	62.97	59.84	60.00
TiO_2	2.67	2.05	1.72	2.47	2.67	2.70	0.62	0.72	0.51
Al_2O_3	15.51	16.30	16.20	17.10	16.10	16.22	15.54	17.63	15.01
Fe_2O_3	4.51	0.70	2.65	8.36	5.11	4.48	6.84	5.87	8.03
FeO	7.47	8.99	6.39	2.28	5.38	7.15	1.86	3.13	1.62
MnO	0.19	0.19	0.20	0.21	0.25	0.21	0.26	0.29	0.50
MgO	5.71	4.54	3.84	2.41	3.26	5.26	0.44	0.43	0.43
CaO	10.82	8.63	7.24	5.49	6.48	10.33	1.33	1.70	1.11
Na_2O	2.54	3.42	3.15	4.54	4.31	2.27	5.34	6.14	4.86
K_2O	0.93	1.96	2.68	2.47	2.62	1.12	5.06	4.83	5.40
P_2O_5	0.71	0.64	0.51	1.24	1.22	0.57	0.12	0.17	0.10
H_2O	1.76	0.52	1.20	1.73	0.82	1.58	1.80	1.64	2.75

1. Average, 41 basalts, subvolcanic basement, Silali volcanics and contemporary flows.
2. Hawaiite, Naudu, Silali volcanics, normative andesine, 61553*.
3. Hawaiite, Kwakoinyangu, Silali volcanics, 61490.
4. Mugearite, Kaparnat, subvolcanic basement, normative oligoclase, 61631.
5. Mugearite, Jebunbun, Silali volcanics (?), 61561.
6. Average, 8 gabbro inclusions in Katenmening basalts, Silali volcanics.
7. Average, 22 trachytes from the subvolcanic basement.
8. Average, 29 trachytes from Silali volcanics.
9. Average, 6 trachyte tuff lavas from the subvolcanic basement.

*Numbers refer to collection of the Geology Department, University of Western Australia.

The development of Silali will be described in detail in a separate publication, as will the petrological and geochemical results. In this account, which must be regarded as a preliminary summary of the geochemical evidence, no detailed maps or tables can be given, but the essential elements of Silali together with its location are shown in Fig.1.

GEOCHEMISTRY

The results of 125 full major element analyses by XRF methods, and analyses by the same methods covering 13 minor elements (Rb, Cu, Sr, Ba, Zn, La, Y, Zr, Pb, Nb, Cr, Co and Ni) for all but four of these specimens, have been compiled and assessed. The rock samples were collected by McCall in field reconnaissances totalling three months, in 1965 and 1967. Preliminary notes have been published (McCall, 1968a,b) and an account published of the gabbroic and syenitic enclaves discovered (McCall, 1970). A coloured map has been prepared of the entire area mapped: details of location of specimens analysed will be published together with this map. Hornung carried out all but the four additional

10	11	12	13	14	15	16	17	18	Factor 14/1 (41 basalts, 51 trachytes)
59.36	62.80	60.80	59.30	61.41	59.10	56.00	66.90	67.90	+ 1.31
0.43	0.59	0.93	0.36	0.67	0.52	0.66	0.64	0.61	−0.243
14.66	15.60	18.40	20.80	16.58	17.10	15.40	13.00	13.10	+ 1.06
8.87	6.73	3.53	0.84	6.36	4.93	5.62	8.58	8.41	+ 1.42
0.41	1.96	2.12	3.13	2.49	4.48	0.44	0.22	NIL	−0.333
0.30	0.20	0.12	0.18	0.28	0.29	0.25	0.13	0.27	+ 1.49
0.33	0.34	1.05	0.46	0.44	0.40	0.57	0.18	0.25	−0.077
1.50	0.50	1.17	0.95	1.51	1.53	1.30	0.33	0.50	−0.138
5.00	6.62	6.26	5.87	5.74	7.43	10.40	3.70	2.87	+ 2.26
5.30	4.71	4.51	6.77	4.95	5.18	5.12	5.30	4.83	+ 5.30
0.16	0.18	0.22	0.14	0.15	0.14	0.22	0.12	0.66	−0.210
3.14	0.81	2.00	0.93	1.72	1.00	3.68	1.48	2.18	–

10. Average, 3 trachyte tuff lavas from Silali volcanics and contemporary flows.
11. Syenite, enclave, Lokitet, subvolcanic basement, 61594.
12. Syenite, enclave, Karmosit, subvolcanic basement, 61664.
13. Syenite, enclave, Arzett, Silali volcanics (Ne normative), 56155.
14. Arithmetical mean of 7 and 8, taken as an average trachyte lava.
15. Phonolitic trachyte, Kuttung, Silali volcanics, 56195.
16. Phonolite (glassy), west of Black Hills, Silali volcanics, 56198.
17. Comendite, Lokitet, subvolcanic basement, 56237.
18. Comendite tuff lava, Lokitet, subvolcanic basement, 56241.

major element analyses, which were done by J. Graham of the C.S.I.R.O., Perth, Western Australia. The analytical results are summarised in Tables I and II.

The geochemical results cover basalts of the foundation beneath Silali, together with trachytes, comendites and syenite enclaves: and basalts, gabbro enclaves, trachytes, phonolitic trachytes, phonolites and syenites of the volcano itself. There is no obvious disparity between the mineralogy, petrography and chemistry of the two suites. A satisfactory division for tabulation purposes is:

(1) basalts, hawaiites and mugearites;
(2) gabbro enclaves in basalts;
(3) trachytes, comendites, phonolitic trachytes and phonolites;
(4) trachyte and comendite tuff lavas and tuffs;
(5) syenite enclaves in trachytes.

Classification of the basaltic rocks is difficult; anomalous are basalts that contain less than 50% silica but have andesine or oligoclase in the CIPW Norm. Of 45 basaltic rocks analysed, 34 are truly basaltic, 6 of this anomalous type, 2 are hawaiites and 3 are

TABLE II

Minor elements

Element	Trachytes and syenites (72 analyses)		Basalts and gabbros (49 analyses)		Factor
	average (p.p.m.)	range (p.p.m.)	average (p.p.m.)	range (p.p.m.)	
Rb	142	55– 260	16	0– 55	+ 9.0
Cu	20	0– 50	73	0– 160	– 0.275
Sr	42	0– 535	617	265– 388	– 0.068
Ba	200	0–2250	884	225–5100	– 0.242
Zn	208	20– 330	85	45– 200	+ 2.4
Y	96	8– 230	34	20– 70	+ 2.8
La	103	20– 210	16	0– 65	+ 6.4
Zr	916	70–1920	237	75–1450	+ 3.9
Pb	15	0– 45	2	0– 15	+ 7.5
Nb	335	80– 410	27	1– 85	+ 12.4
Cr	54	10– 190	155	50– 480	– 0.347
Co	44	0– 155	71	0– 180	– 0.620
Ni	321	65–1270	509	165– 915	– 0.631

K/Rb ratio: basalts = 580; trachytes etc. = 495.

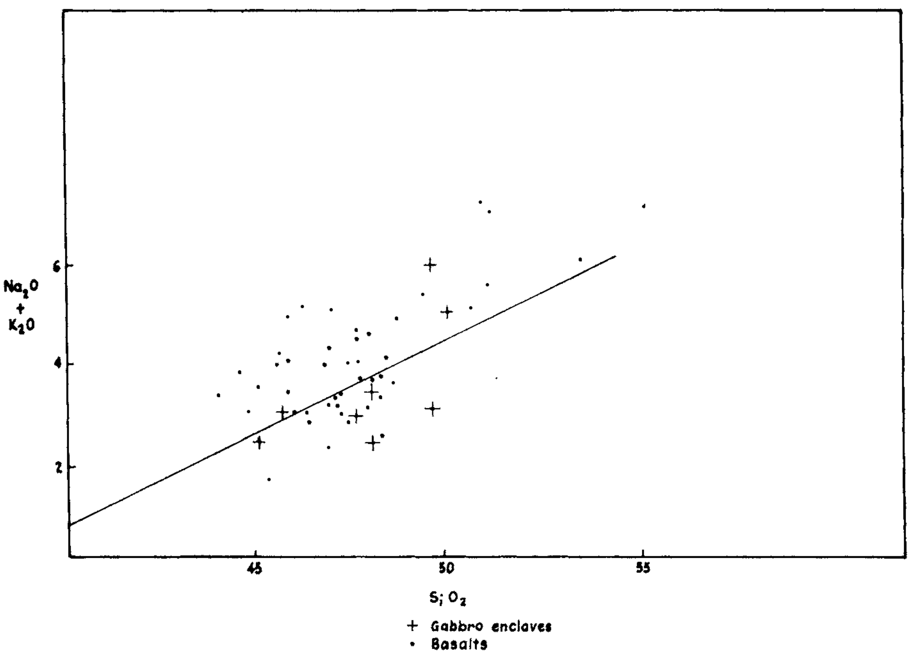

Fig.2. Silali and foundation. Alkali/silica diagram – basalts and gabbros.

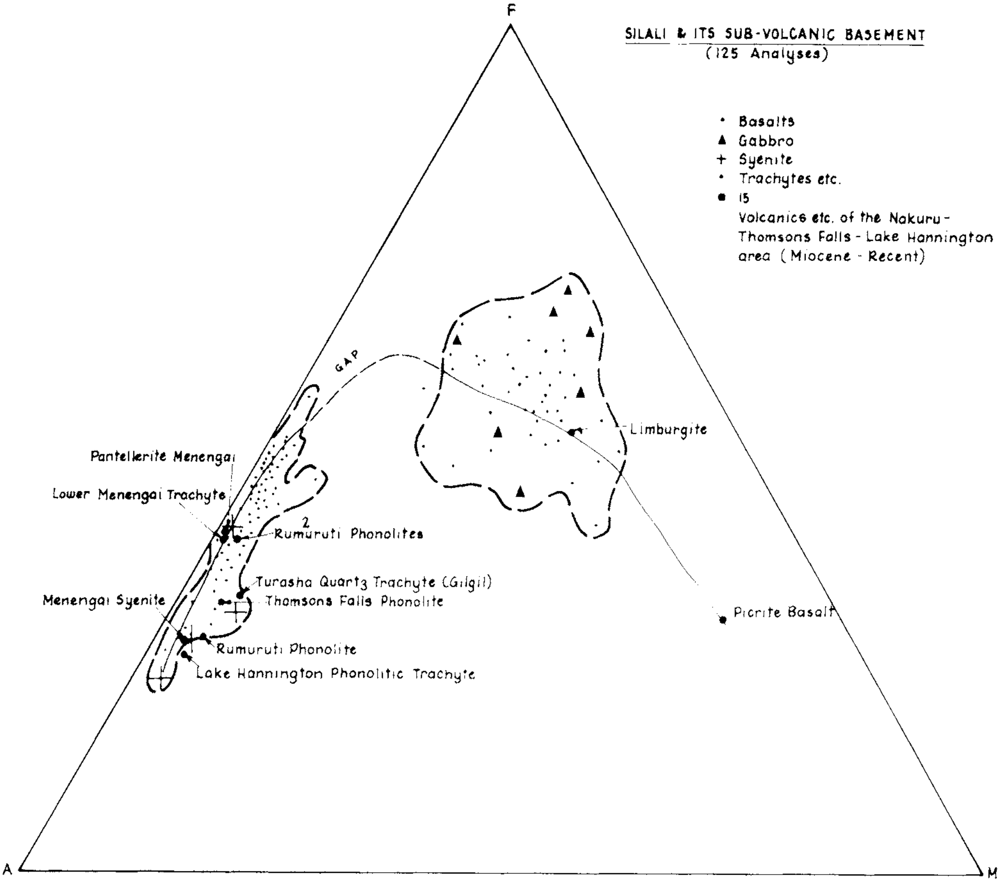

Fig.3. Silali and foundation. FMA diagram. The line shows the trend of differentiation. Fifteen analyses of Miocene–Recent rocks of the Nakuru–Thomson's Falls–Lake Hannington area are plotted for comparison (McCall, 1967).

mugearites (the hawaiites and mugearites containing between 50% and 53% silica in addition to markedly increased alkali content). Of the basalts, quartz, olivine and nepheline normative examples are represented. The mugearites are abnormally oxidised, ferric iron predominating. Of the eight gabbro enclaves analysed, one is of the anomalous type noted above.

Of 65 intermediate to acid volcanic rocks analysed, 51 are trachytes, one a hyaline phonolite, one a comendite, one a phonolitic trachyte, 9 trachyte tuff lavas, one a comendite tuff lava, and one a trachyte tuff. Three syenite enclaves were analysed.

The tuff lavas are abundant in the foundation sequence, and are found to have a high normative quartz content, and potash in excess of soda in the analysis, the inverse of the case in most of the trachyte lavas. Similar tuff lavas are known in the Silali sequence, but are restricted to thin surface veneers to normal flows (cf. McCall, 1964). Such occurrences have been noted in small spills of trachyte down the flanks of satellite cones, occurrences

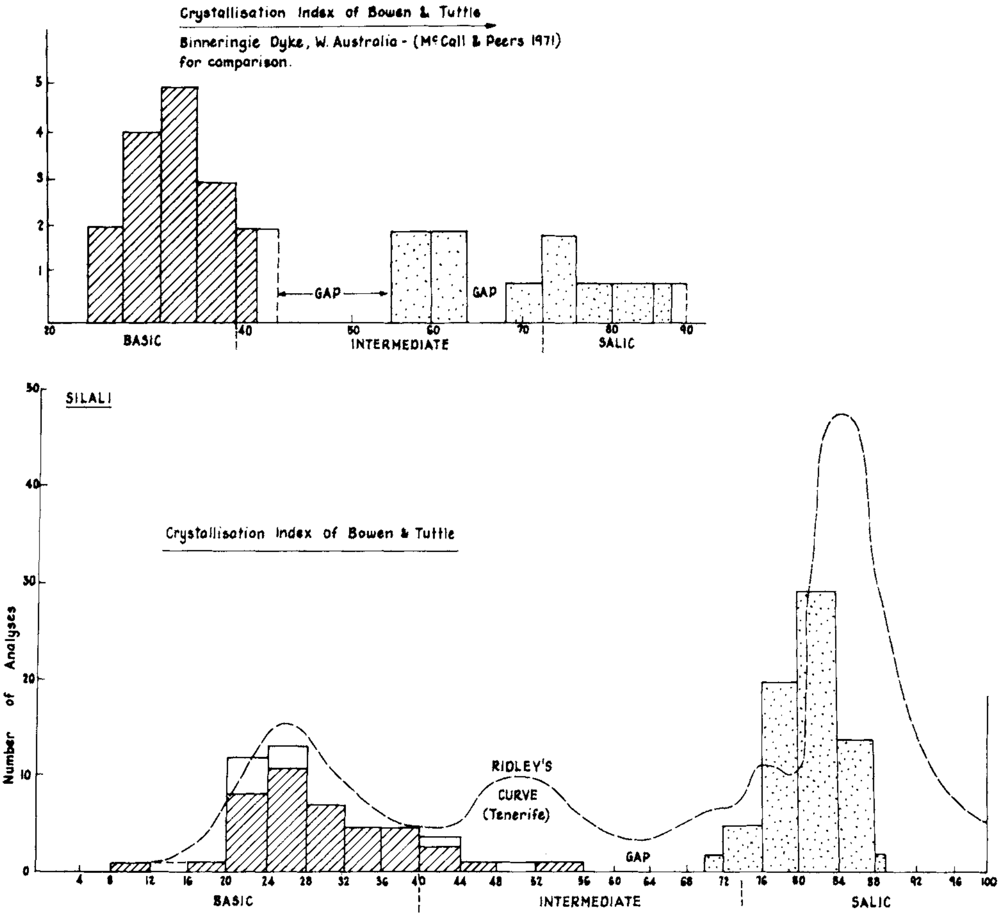

Fig.4. Crystallisation index–silica content diagrams. Silali and foundation (below); Binneringie Dyke (for comparison, above); Tenerife (dotted curve below, for comparison).

that can only be reconciled with surficial autoclasis in a lava flow or sintering of ejecta falling onto it.

In Fig.2 and 3 alkali–silica and FMA plots are given. Crystallisation index statistics are compared in Fig.4 with those of Ridley for Tenerife (1970), and McCall and Peers for the Binneringie Dyke, Western Australia (1971) (see discussion section). Silica contents are compared with the latter in Fig.5 and Fig.6 shows the major element variation trends, evident weakly in the hawaiites and mugearites; and fully developed across the gap of bimodality in the intermediate to acid rocks.

DISCUSSION

Ridley (1970) documents a statistical discontinuity in the basalt/trachyte suite of Tenerife; though he urges caution in the use of statistical abundances in total samplings

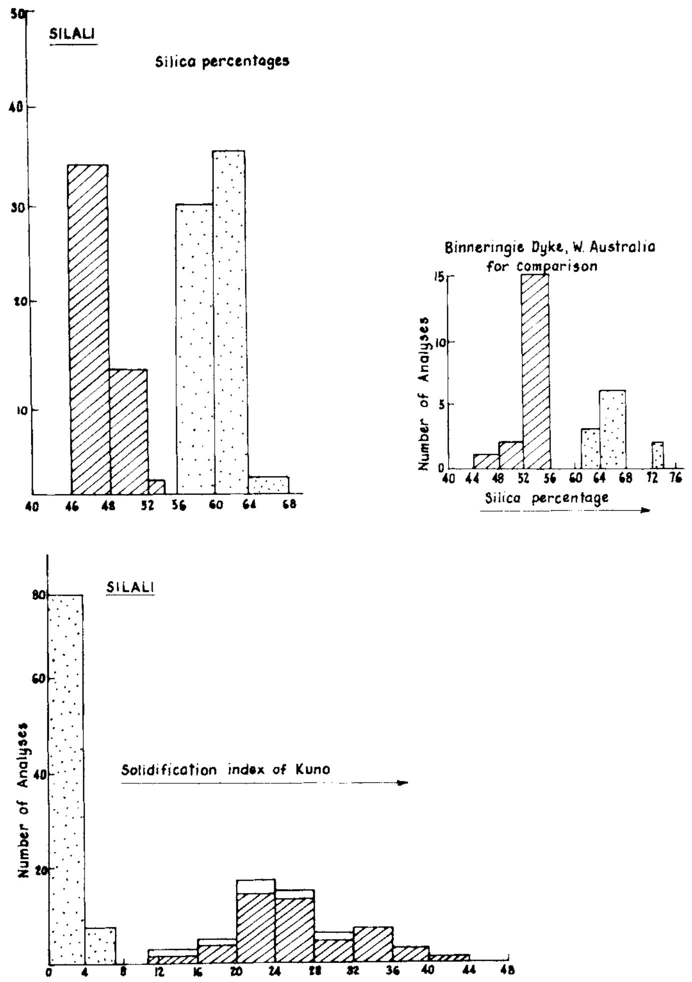

Fig.5. Silica contents: Silali and foundation (left, above); Binneringie Dyke (for comparison, right, above); Solidification indices: Silali and foundation (below).

from subaerial volcanics, the evidence from Silali, covering both volcanic and intrusive rocks, indicates that:

(1) The bimodality is real.

(2) It cannot be simply explained by disproportional sampling.

(3) It applies to volcanic rocks and their intrusive equivalents, gabbro and syenite enclaves that plot exactly with the volcanic rocks on the FMA and silica/alkali diagram.

(4) A distinct gap replaces the tenuous link on the crystallisation index/abundance curve of Ridley.

(5) The rocks of the foundation beneath Silali (of Plio-Pleistocene age), of the Quaternary volcano itself, and the fifteen rocks analysed previously from the Nakuru–Thomson's Falls–Lake Hannington Area (McCall, 1967), all show the same bimodality. The latter

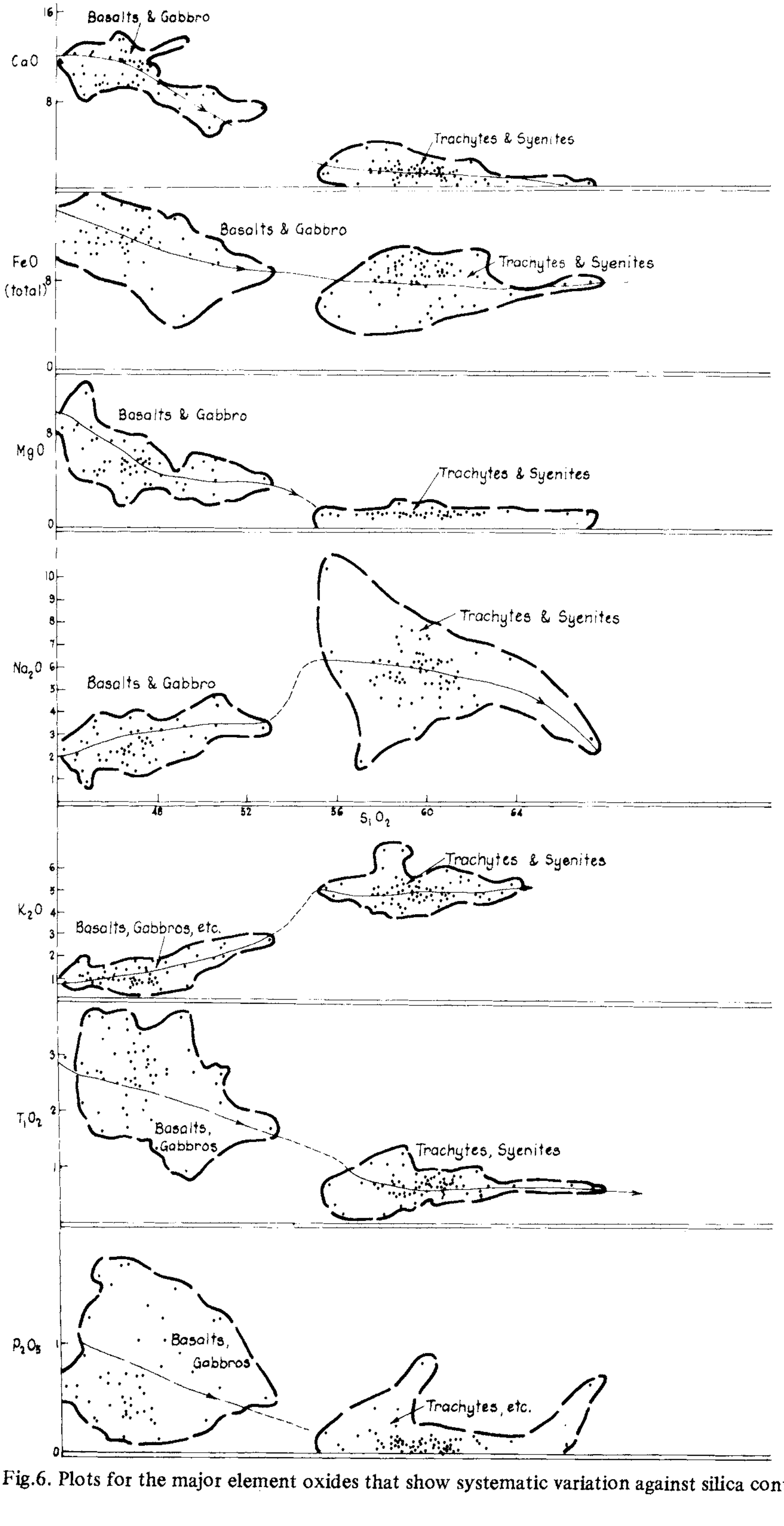

Fig.6. Plots for the major element oxides that show systematic variation against silica contents.

group include Miocene basalts and phonolites, and these show no departure at all from the trends and fields of the Silali plots for later eruptives, a fact that confirms the suggestion that this bimodality is fundamental to the Tertiary–Quaternary eruptivity of the central Gregory Rift Valley, and that one genetic line of descent must cover the earlier phonolites and later trachytes, quartz–trachytes and comendites. In this connection it is significant to note that there are minor occurrences of phonolite in the later suites, just as there are minor occurrences of trachytes, etc., in the Miocene suites.

There is a history of argument concerning supposed bimodality of *oceanic* basalt trachyte associations (Chayes, 1963). Bimodality is not evident in Hawaii (MacDonald, 1963), Tristan (Harris, 1963) or St. Helena (Baker, 1968). Ridley attributes the bimodality in Tenerife to the abundance of salic material in the later eruptivity, short explosive eruptions only sampling part of the magma reservoir or salic material remaining available in reservoirs for long periods of time. Baker and McReath (1971) remark, in the context of salic rocks of the ocean basins, that although a scarcity of intermediate lavas has been cited as an obstacle to fractional crystallisation, the meaning and even the existence of such a gap have become increasingly doubtful. Even so, they accept the need to invoke gas transfer to explain anomalous residual end members, adding that strontium isotope studies have cast doubt on the concept that a straightforward fractional crystallisation sequence links parental basalts to derivative trachytes, phonolites and rhyolites. Differentiation must, at the least, be now viewed as a more complex process involving intermediate stages of solidification and refusion. They reject sialic remelting as the origin for the *oceanic* intermediate rocks associated with basalt volcanism, and imply a mantle provenance for them, albeit involving complex differentiation processes.

The evidence from Silali establishes that bimodality is real, and it must be a characteristic that may be developed in both *oceanic* and *continental* alkalic basalt suites, but is not necessarily developed. The intermediate/acid rocks of the Silali *continental* suite must be related, as Baker and McReath suggest, to complex differentiation from a basalt stem, itself a partial melting product at upper mantle level. How else could the minor element contents (Cr, Co, Ni) be so similar? Neither the major element composition of the intermediate to acid rocks, so markedly peralkaline and so uniform, nor the minor element composition, is compatible with an origin in sialic remelting (McCall, 1968b).

The minor element contents like the major element contents show definite trends, incipient in the mugearites and hawaiites, and fully developed in the intermediate/acid rocks.

In a geochemical study of Paka volcano to the south of Silali, by Sceal and Weaver (1971) the points are made that zircon does not enter major crystalline phases (except possibly early crystallising pyroxene), and is, therefore, a residual element, remaining in the liquid phase in any liquid solid reaction or equilibrium (Chao and Fleischer, 1960; Taylor, 1965). Rb, Nb, La and Y behave similarly. Zirconium has an enhanced solubility in peralkaline liquids.

Both the increase in these elements and the decrease in Sr and Ba strongly indicate prolonged fractionation as the sources of the trachytes and comendites: barium and

strontium contents in the melt being used up largely in the early felspar separation, Sr to plagioclase and barium to alkali felspar marked. Depletion in strontium (in the case of Silali × 0.068) can only be produced by prolonged fractionation (Noble et al., 1969).

The zirconium, niobium and rare earth data indicate substantial increases, calling to mind the similar increases in carbonatites, but unlike the carbonatites there is a decrease in Cu, Sr and Ba in these intermediate/acid differentiates. Despite this difference, a comparison with the carbonatites is not out of place, for these intermediate/acid rocks display the mineralogy and mineral paragenesis of the agpaites (McCall, 1957c, 1968b). This is a most important observation, for it suggests that these intermediate/acid differentiates may be products of an agpaitic differentiation process, typical of a cupola environment. In such a differentiation, controlled by excessive upward movement and concentration in the cupola of volatiles, felspars and felspathoids crystallise out first, and ferromagnesians of the acmite–aegirine–fayalite–kataphorite–cossyrite–riebeckite–arfvedsonite assemblages crystallise late: this is a reversal of the Bowen-type fractionation sequence, in which ferromagnesians crystallise out first. It is suggested that high and low level reservoirs existed beneath the central Gregory Rift Valley; all the field evidence suggests that basalts were erupted from deeply penetrative fracture systems and trachytes from high standing cupola reservoirs. Whether the two reservoirs were finitely separated in space or simply bounded by an abrupt interface (of immiscibility?) cannot be determined on the basis of the evidence at present available, but differentiation appears to have taken place quite separately under two quite different regimes, the high level reservoirs coming under a regime of concentrated volatiles.

Whether we should regard the processes involved in these high level reservoirs as metasomatic (McCall, 1957c) or simply as involving *crystal–liquid–alkali charged vapour* systems (MacDonald et al., 1970) is immaterial. Exactly how the enrichment in silica and alkalies, minor impoverishment in total iron and oxidation of the iron, is effected remains unknown, and is a subject for experimental and theoretical study. Certainly there is no smooth progression from basic to acid magmas, and the processes involved must be complex.

The deep intracrustal magma reservoirs are revealed to us in older, dissected geological systems, and here we might look for a clue to the nature of this magma reservoir dichotomy. We do, in fact, see in such environments batholiths, stocks and large layered intrusions that are congealed magma reservoirs, and these commonly reveal abrupt interfaces between successively congealed phases of contrasting mineralogy and chemistry. Examples are layered intrusions containing abruptly defined and hiatally separated (chemically and mineralogically) peridotite, pyroxenite, gabbro and granophyre phases: the abruptly bounded contrasting phases of the Peruvian batholiths (studies by W.S. Pitcher, as yet unpublished): or the alkaline ring complexes of Africa. Specifically, we can draw a useful comparison with the Binneringie Dyke (320 × 3 km differentiated major intrusion of dyke form in Western Australia: McCall and Peers, 1971) and the Silali evidence. In that case we can recognise bimodality in a congealed, differentiated magma chamber. The abruptly defined phases of basic and intermediate/acid rocks could represent, on a minor scale, the

sort of relationships beneath the Rift Valley, where, however, the intermediate/acid reservoir must have been to the same order of volume as the basic reservoir. In the Binneringie Dyke, intermediate rocks between 56% and 61% silica content are completely absent, yet this is clearly a differentiation sequence: fractionation was not smoothly progressive. Such bimodality can only be a product of complexities in the differentiation itself, and thus the comparison with Ridley's Tenerife histograms and the Silali histograms in Fig.5 and 6 is highly significant. It is also significant that there is abundant evidence that complications were introduced by residual concentrations of volatiles, in the case of the Binneringie Dyke.

There is, in the suites studied from Silali, a range of compositions from intermediate to acid composition, from 56% to 66% silica content, and from 13.10% normative nepheline and 4.38% normative nosean at one extreme to 31.67% normative quartz at the other. Despite this wide range, all these rocks must be regarded as differentiation products derived from a separate, high level magma reservoir environment to that which supplied the basaltic–hawaiitic–mugearitic lava flows.

Other unanswered questions remain. Does liquid immiscibility have some bearing on this separation of discrete magma reservoirs? What is the explanation for the volume anomaly, and can ease of access to the surface of the cupola-derived magmas fully explain it, or must it be referred back to the anomalous differentiation process, whatever its nature? Has the explanation of McBirney (1968) for basalt–andesite bimodality involving overthrusting and the rise of water into an overthrust block any relevance to the Rift Valley? The graben-in-graben geotectonic setting of Silali is quite devoid of any suggestion of thrust tectonics, though thrusts could conceivably exist in depth; the lack of water availability in the underlying granite–gneiss basement is another objection to any such derivation. The intermingling of the two eruptive suites, intimately in time and space, must be considered in any theorising, and is strongly against such a concept, against sialic remelting or contamination, and against partial melting at two different levels within the upper mantle.

CONCLUSIONS

The intermediate/acid eruptives of the central Gregory Rift Valley in the Silali area may be taken as a model for eruptivity there extending from Miocene times to the present day. They must be derived by complex differentiation processes, operating in discrete magma reservoirs quite separate from the basaltic reservoirs and situated high in the crust beneath the Rift Valley, in the form of cupolas. The differentiation process is an extension of the agpaitic pattern, but the parental material was basaltic, not nephelinitic. The geochemical relationships, for major and minor elements, leave little doubt that these are differentiates of the basalts themselves of upper mantle provenance, if hiatally separated from them, not products of sialic remelting or sialic assimilation. The reason for the bimodality and the separation of discrete differentiating magma reservoirs is not understood, but must be related to cupola concentration of volatiles (gas transfer). It is evident that

the relative abundance of such intermediate/acid differentiates in any sector of the Great Rift Valley cannot be taken as a direct indication of the thickness of the underlying sialic crust.

ACKNOWLEDGEMENT

The author is indebted to J. Sceal for sending him a copy of his report on similar studies of Paka volcano immediately south of Silali, and to Dr. L.A.J. Williams of Nairobi University College and Prof. Gerasimovsky of the Vernadskii Institute, Moscow, for discussions of these results.

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THE SEISMICITY OF THE EAST AFRICAN RIFT SYSTEM

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ABSTRACT

Fairhead, J.D. and Girdler, R.W., 1972. The seismicity of the East African rift system. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 115–122.

Epicentres for the whole of Africa for the period 1963–1970 have been relocated (Fig.1) and their relationship to the rifting studied. For the Gulf of Aden and Red Sea most of the epicentres are found to be related to the spreading zones along their axes. In East Africa, the epicentres appear scattered but are found to be closely associated with the recent rift faulting in the ancient Precambrian crust. Earthquake mechanism studies show the Gulf of Aden and Red Sea to be associated with tensional stress fields directed north easterly and the rifting in East Africa to be related to a tensional stress field in an east south easterly direction. By looking at ray paths from various groups of earthquakes to various African recording stations, an attempt has been made to map the region of P slowing down associated with the rifting (Fig.2). The P wave delays are related to the negative Bouguer gravity anomalies and a possible model for the structure of the lithosphere is presented (Fig.3).

INTRODUCTION

The seismicity of Africa has been studied for the period 1963–1970. The epicentres were relocated using the Joint Epicentre Method (Jed) of Douglas (1967), the large earthquake occurring in northern Tanzania on March 7, 1964 being used as the master event. The 26 largest earthquakes which occurred during this period were relocated with respect to this master event and some of these were restrained as submasters against which all the other events were relocated. The object of this exercise was to obtain the best possible location of epicentres for comparison with the rift faulting and local geology. In addition, all the largest earthquakes were examined for possible fault plane solutions to aid the understanding of the stress fields associated with various parts of the rift system. The Jed method of relocating epicentres also gives estimates of travel time corrections and these are used in an attempt to map the region of slowing down of P velocities beneath regions of rifting.

Results of these studies have been given in considerable detail by Fairhead and Girdler (1970, 1971) and these two papers contain references to previous works. Only the main results are given here with special emphasis on those which give information on the structure and evolution of the rift system. The work on P wave delays is combined with

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previous gravity studies (Girdler et al., 1969; Girdler and Sowerbutts, 1970; Searle, 1970) in an effort to gain a picture of the deeper structures and nature of the upper mantle associated with the rift system.

SEISMICITY

The relocated epicentres for the period 1963 through 1970 are shown in Fig.1. To a first order, the East African rift system may be considered as three plate boundaries joining in the Afar region of Ethiopia. These are the Gulf of Aden, the Red Sea and the Gregory Rift extending through Ethiopia, Kenya and Tanzania to Lake Nyasa. A branch (the Western Rift) contains Lakes Tanganyika, Kivu and Albert. The developing plate boundaries are considered in turn.

The Gulf of Aden

It is seen from Fig.1 that the epicentres in the Gulf of Aden are confined to its axis. The Gulf of Aden is oceanic and there is a sea floor spreading centre along its axis (Laughton et al., 1970). When the epicentres are examined in detail (see for example, fig. 11 of Fairhead and Girdler, 1970) they are found to be associated either with the axial spreading zone or with northeasterly transforms.

The Red Sea

In the Red Sea, most but not all of the epicentres are confined to the deep axial trough. Like the Gulf of Aden, the Red Sea also has a sea floor spreading centre (Vine, 1966) but so far transform faults, although suspected, have not been mapped. The group of earthquakes in the north are the foreshocks and aftershocks associated with the 1969 March 31 earthquake in the mouth of the Gulf of Suez. In the southern Red Sea, some epicentres are associated with the western margin (Fig.1); these seem to be related to the rotation of the Danakil horst and the creation of two spreading centres in the south.

East Africa

In East Africa the epicentres on first sight seem to show much scatter. When plotted on maps showing geological faulting (fig.5, 6 and 9 of Fairhead and Girdler, 1971) it is seen that there is a good correlation between the epicentre locations and recent faulting. The scatter appears to be due to the long history and complex nature of the continental crust. The contrast between the seismicity of East Africa and that of the Gulf of Aden and Red Sea seems to be related to the degree of evolution of the plate boundaries. In East Africa the continental crust has not broken apart and the epicentres are located in a crust with a history of thousands of millions of years whereas in the Gulf of Aden where the plates are separating, the oceanic crust has a history of but a few million years.

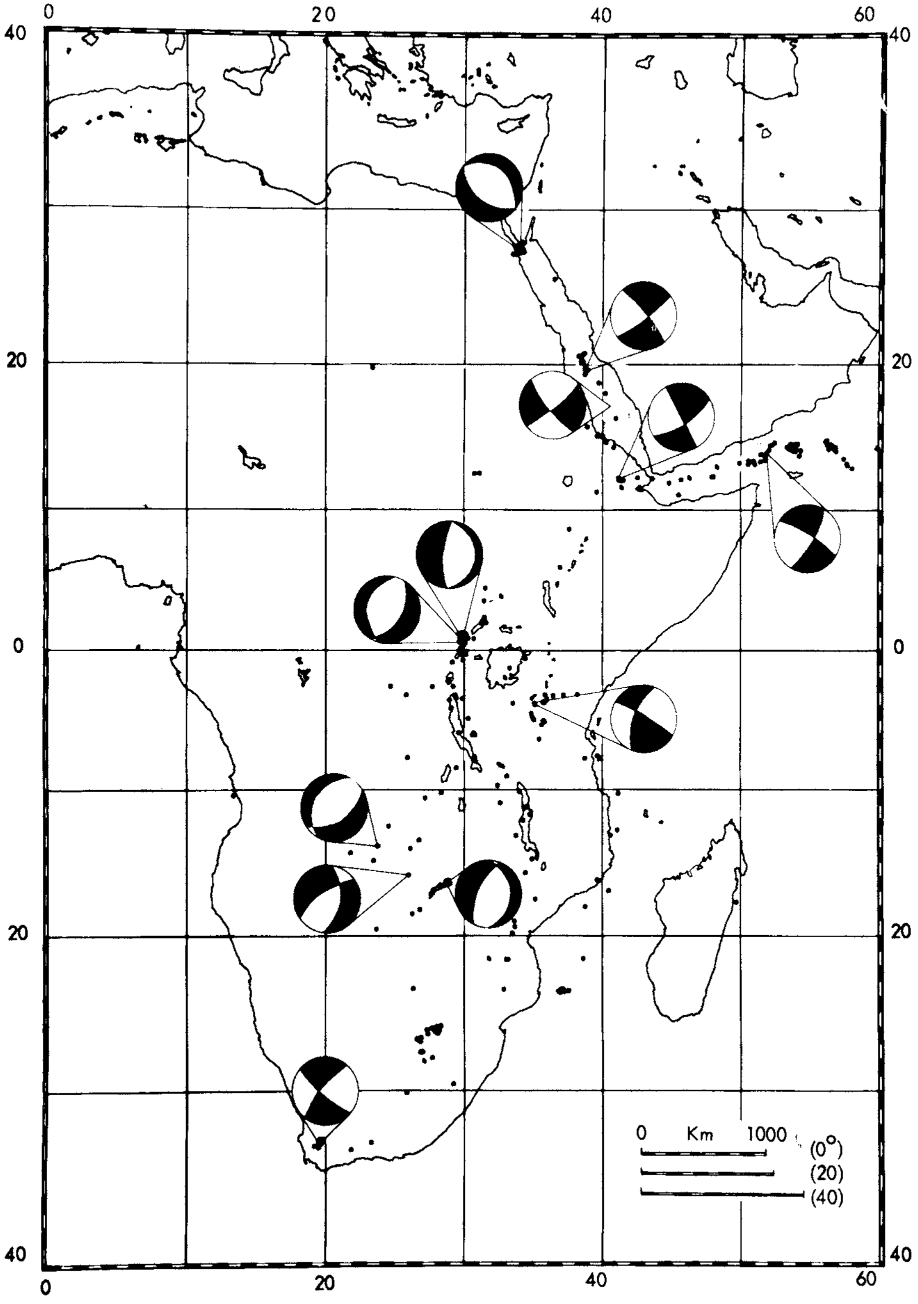

Fig.1. Map showing the seismicity of Africa (1963–1970) and fault plane solutions (equal area projections of the lower hemisphere of the focal sphere) to the end of 1970 (by courtesy of the Royal Astronomical Society).

EARTHQUAKE MECHANISMS

Fourteen fault plane solutions are now available and twelve of these are shown in Fig. 1. Of the two omitted, one (1969 March 20) is in Ethiopia and belongs to the Sardo sequence and has the same mechanism and almost the same location as the event of April 5, 1969; the other (1963 September 25) is in Zambia and has the same mechanism and almost the same location as the event of September 23, 1963.

It is seen from Fig.1 that the fault plane solutions indicate either strike-slip or normal fault mechanisms. There is no evidence of any compression. Sometimes, where the geology is not well known, it is not easy to choose the fault and auxiliary planes. The most likely interpretations are shown in Fig.2. For the Gulf of Aden and Red Sea, the stress fields are northeasterly, i.e., consistent with the separation of the Arabia-Somalia and Arabia-Nubia plates about poles of rotation in northern Africa. In East Africa, the situation is not so clear but the tensional stress field seems to be directed in a direction about N 120°. This suggests a pole of rotation for Nubia-Somalia somewhere southwest of Africa.

TRAVEL-TIME DELAYS

A study has been made of the travel-time corrections for the seismic recording stations throughout Africa. The corrections are azimuthally dependent. Using Jed, corrections are estimated for various groups of earthquakes with respect to each station. In this way it is possible to study the travel-time corrections for various ray paths crossing regions of rifting and crossing the shield (fig.14 and 15 of Fairhead and Girdler, 1971). It is found that for paths crossing the rift system the travel-time corrections are positive (by up to 3.87 sec) and for paths crossing the shield the corrections are negative. This means that rays travelling near or beneath the rift system are slowed. The effect is particularly pronounced for the Gulf of Aden, for the Red Sea and for the rift in East Africa north of 5°S. Gumper and Pomeroy (1970) observed a similar effect for their studies of P_n, S_n, and Lg phases in Africa. By putting all these results together, it is possible to tentatively map the region of P slowing down associated with the rift system. This is shown by the stippled region in Fig.2.

TRAVEL TIME DELAYS AND GRAVITY ANOMALIES

It is of interest to enquire into the nature of the stippled region shown in Fig.2. From the seismic delays, it can be inferred that there is lower velocity material beneath the rift zone. Gumper and Pomeroy (1970) from the attenuation of S_n infer the existence of a gap in the mantle portion of the lithosphere beneath the northern part of the rift system. Girdler et al. (1969) independently arrived at a similar inference from a study of the long wavelength negative Bouguer anomaly over East Africa. The negative anomaly dies out at

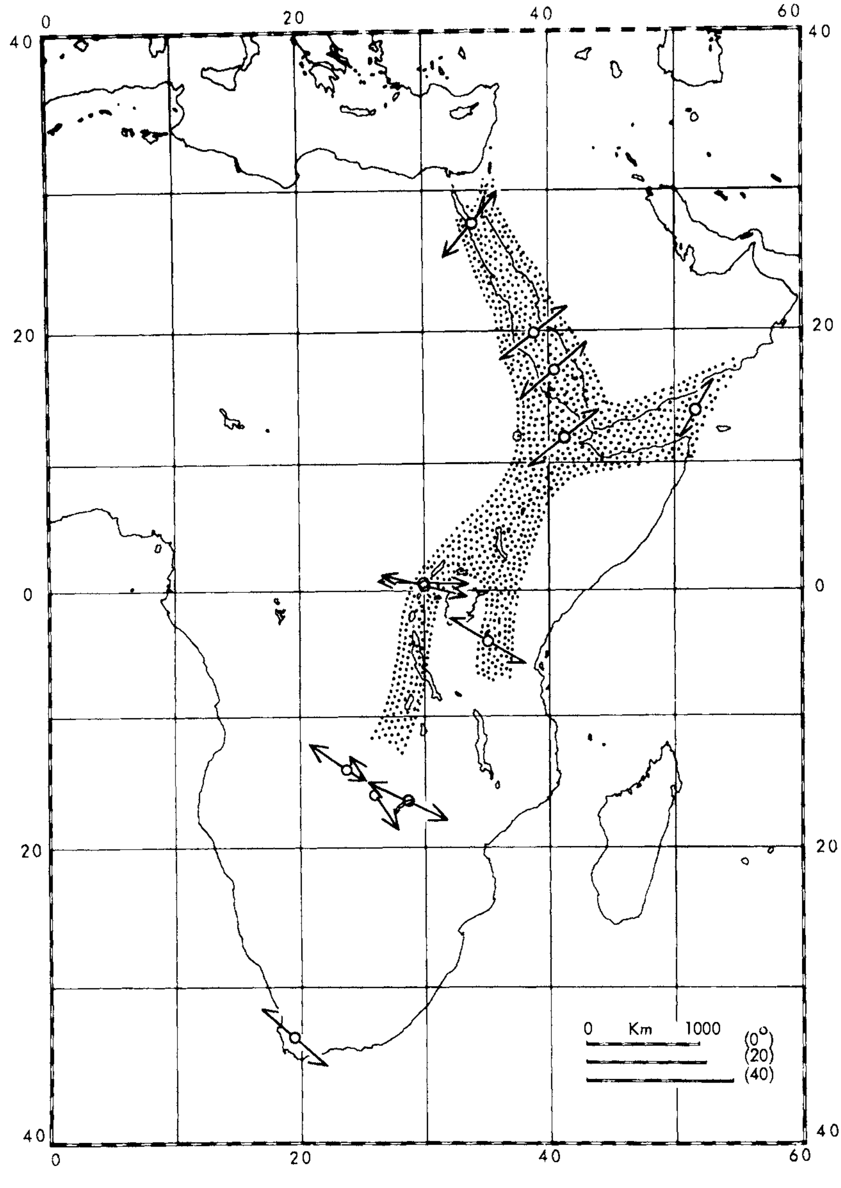

Fig.2. Map showing possible stress fields with which various parts of the rift system may be associated and tentative mapping of the extent of the region of slow P propagation and thinning of the lithosphere (stippled) (by courtesy of the Royal Astronomical Society).

about 5°S (Girdler and Sowerbutts, 1970) and this is about where the travel time anomalies die out and there is a change in the nature of the rifting from a complex rift valley with grid faulting and volcanism (to the north) to a more simple region of block faulting (to the south). It seems that the region of P slowing down, S_n attenuation and negative Bouguer anomaly are all related.

The nature of the gravity field across the rift zone in East Africa is now well established. A typical Bouguer profile north of 5°S shows a long wavelength negative anomaly with a superimposed smaller positive anomaly over the rift axis. This type of profile is typical of

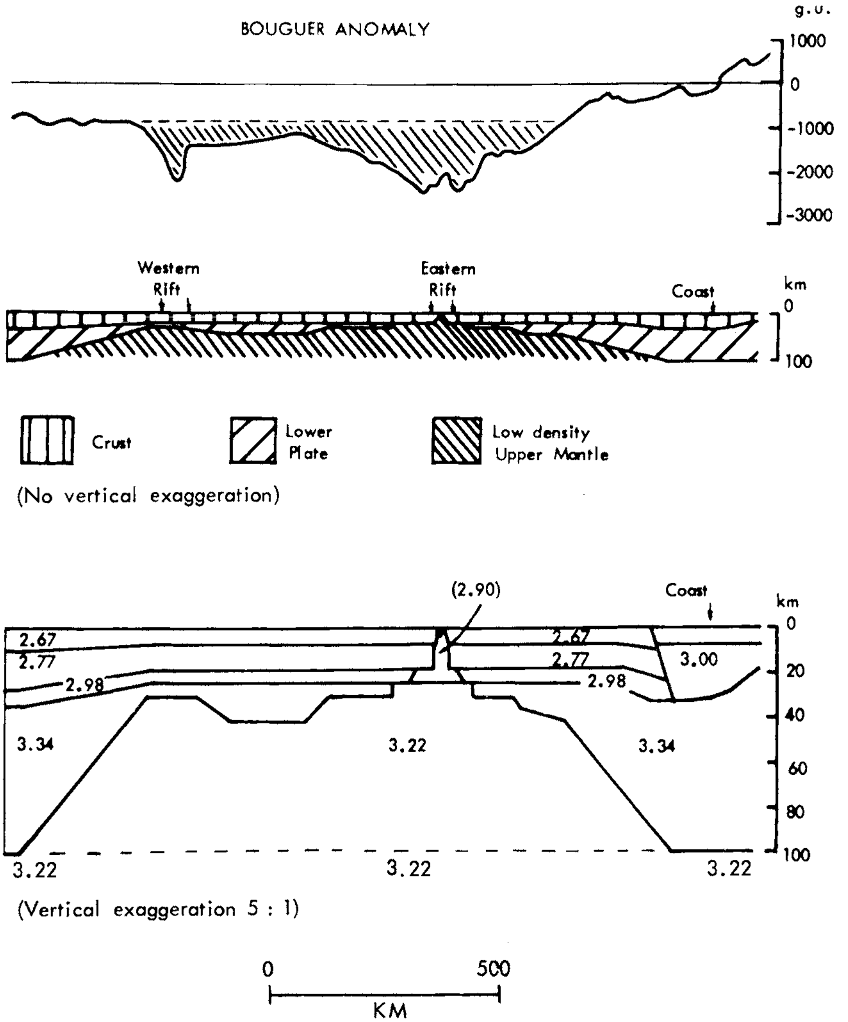

Fig.3. The Bouguer gravity anomaly over the rift zone in the neighbourhood of the equator. The model (one possible interpretation) shows thinning of the lithosphere (crust and upper mantle) and emplacement of lower density asthenosphere. Details of the model including specific gravities are shown in the lower diagram.

the Gregory Rift in Kenya and Ethiopia and the Red Sea and Gulf of Aden.

In East Africa, the negative anomaly has a width of 1000 km and an amplitude of up to -150 mgal. The gradients are small suggesting the light material causing the anomaly is at considerable depth and has small density contrast. The anomaly has been interpreted by Girdler et al. (1969), Girdler and Sowerbutts (1970) and Baker and Wohlenberg (1971) as being due to lower density asthenosphere replacing the upper mantle part of the lithosphere (specific gravity contrast, -0.12).

The axial positive anomaly has been mapped in considerable detail between 0.25°N and 1.25°S by Searle (1969, 1970). Searle made 1550 gravity measurements over the rift floor and shoulders and found the anomaly to be 40 to 80 km wide with an amplitude of +30 to +60 mgal. His most favoured interpretation suggests the presence of a mantle-derived intrusive zone about 20 km wide and in places reaching to within 2 km of the rift floor. There is thus extreme thinning of the lithosphere beneath the rift accounting for the volcanicity.

The development of the positive anomaly is dependent on the extent to which the axial intrusive zone has developed. Over the Gulf of Aden and Red Sea, the positive anomaly is much larger (e.g., reaching +150 mgal over parts of the Red Sea) and here the intrusive zones are much larger, reaching the surface (Girdler, 1958) with complete separation of the continental lithosphere.

Fig.3 (adapted from Searle, 1969) shows a gravity profile and interpretation illustrating the possible structure of the East African rift zone near the equator. The model assumes thinning of the lithosphere. The subsequent expansion of the asthenosphere into the lower (upper mantle) part of the lithosphere gives the negative specific gravity contrast (-0.12) causing the long wavelength negative anomaly and the further expansion of the asthenosphere into higher levels of the crust gives the positive specific gravity contrasts (+0.13 to +0.23) causing the smaller positive anomaly. It is seen that such a model with a large volume of low density material replacing the upper mantle part of the plate can readily explain the travel time delays associated with the rifting.

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STUDY OF MICROEARTHQUAKES IN THE RIFT ZONES OF EAST AFRICA

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(Received February 1, 1972)

ABSTRACT

Rykounov, L.N., Sedov, V.V., Savrina, L.A. and Bourmin, V. Ju., 1972. Study of microearthquakes in the rift zones of East Africa. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15 (1/2): 123–130.

The field work of recording micro-earthquakes in the southern part of the Gregory Rift and the northern part of the Western Rift (Lake George, Ruwenzori and Lake Albert) was carried out in the months of June, July and August 1968 and 1969.

Data were recorded by a total of five portable automatic seismographs. The network was moved from place to place through the area. Duration of the recording at each selected site was from 5 to 10 days.

Analysis of the data has given new and more precise information about epicentres and depths of earthquakes with magnitudes $1 \leqslant M \leqslant 3$ in the areas under investigation.

Presence of reflected and refracted waves on the earthquake records have given some new data about the structure of the earth's crust in the southern part of the Gregory Rift. The thickness of the earth's crust is 35–37 km; the velocity of P waves is 5.8 ± 0.3 km/sec for the upper layer (thickness about 18 km) and 6.5 ∓ 0.3 km/sec for the intermediate layer.

INTRODUCTION

This report consists of some results of a study of the microearthquakes in different parts of the East African rift system. The field data were obtained during the Soviet East African Expedition in the summers of 1968 and 1969. The seismological group of the expedition consisted of two workers of Moscow University.

The observations were aimed at collecting seismological material by means of a network of autonomous seismic stations installed successively in the most typical areas of the regions under investigation. The basic tasks were:

(1) The use of highly sensitive mobile instruments to single out the most interesting focal zones for a short time study and estimate the possible connections between location of weak earthquakes and specific tectonic structures.

(2) To estimate the energy range of recorded earthquakes and to find if possible some pecularities of their repetition.

(3) To get some information about deep structures of specific parts of the rift system

using the records of the earthquakes obtained by network of seismic stations.

LOCATIONS AND TECHNIQUES

The regions for investigation were chosen on the basis of previous studies of the seismicity of East Africa (Gutenberg and Richter, 1949; Sykes and Landisman, 1964; Wohlenberg, 1967; De Bremaecker, 1959; Drake, 1969). Such selected regions were the southern part of Gregory Rift (from Lake Magadi in Kenya to Mount Hanang in Tanzania) and the northern part of the western rift between Lakes Edward and Albert and including the Ruwenzori Massif.

To achieve these tasks four portable autonomous seismic stations were used in 1968 and five in 1969; the seismographs were modified ocean bottom instruments of Moscow University (Rykounov and Sedov, 1967; Rykounov, 1969). The range of frequencies was 2 – 20 Hz. The recording time without servicing was about 7 days. Each instrument was supplied with quartz chronometers with accuracy 10^{-6}.

The method consisted of moving polygonal networks from place to place. Changes of geometry of the network permitted increased accuracy in the location of the earthquakes.

With networks with distances between seismographs of 30 km, it was possible to record earthquakes with energy less than 10^{10} J. Taking into account the increasing frequency of smaller earthquakes it was possible to get sufficient information about the seismicity of selected regions in a relatively short time.

The distribution of seismographs in the rift zones of East Africa is shown in Fig. 1A.

RESULTS

(1) The southern part of the Gregory Rift was investigated using ten polygons. The seismicity of Ngorongoro, Oldonyo Lengai, Lake Manyara, at the southern end of the rift near the border of the Masai Steppe (Hanang and Kwaraha mountains) were found to be relatively high. The number of events with M between 0.5 and 3.5 was about 30 per day. The seismicity of the mountains Kilimanjaro and Meru was less. Fig. 1B shows the distribution of epicentres; the accuracy of location being about 2 km

It is possible to see that some of the epicentres are connected with fault zones but there are also epicentres which are not connected with any surface geological features (e.g., the southern part of Lake Manyara). The first information about depths of foci was obtained in this region. Fig. 2A shows the vertical cross section which crosses the rift from east to west through Lake Manyara. It is easy to see that it is not possible to determine any focal planes here. The zone of destruction is probably wide and is not connected with any external features of the rift (e.g., with the main escarpment). Fig. 2B shows the vertical cross section along the rift.

In the western branch of the rift system four polygons were used covering Lakes Edward and Albert and the Ruwenzori Massif.

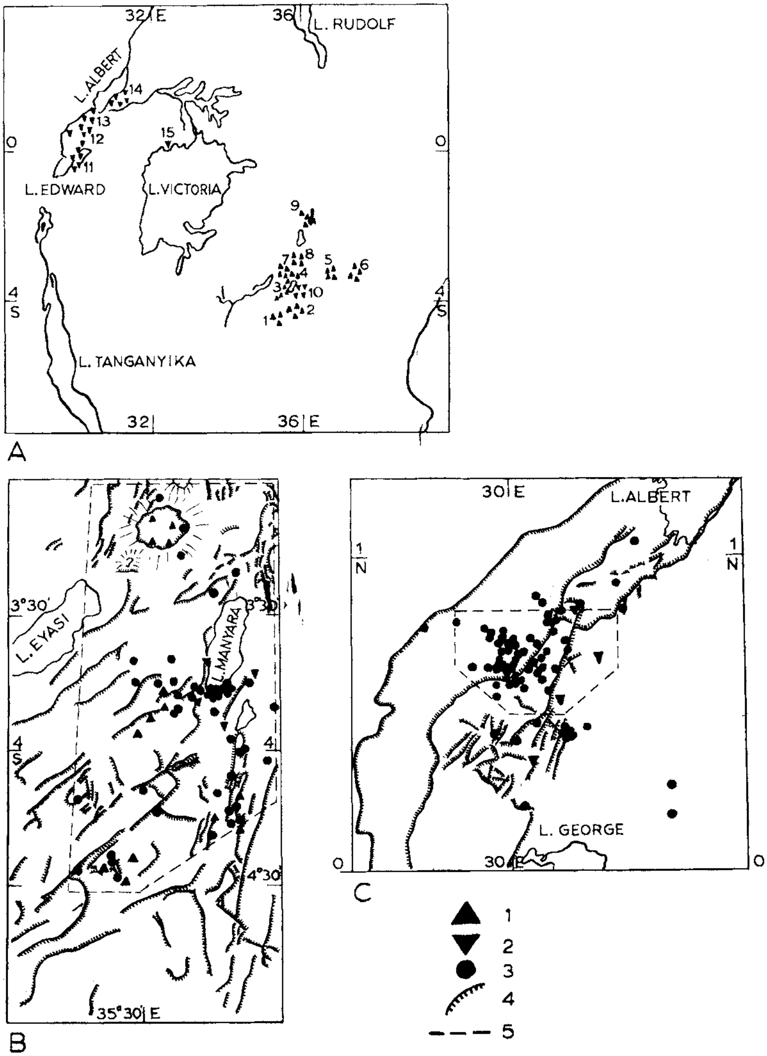

Fig. 1. Distribution of seismographs and epicentres in the regions under investigation. A. Distribution of seismographs and polygons. B. Distribution of epicentres in the southern part of Gregory Rift. C. Distribution of epicentres in the northern part of Ruwenzori. *1* = position of seismographs in 1968; *2* = position of seismographs in 1969; *3* = epicentres; *4* = tectonic lines; *5* = areas of earthquake - recording for estimation of seismicity.

The most seismically active region was Ruwenzori (especially its northern part near the contact with the Lake Albert rift. Fig. 1C shows the distribution of the epicentres in this area.

It is interesting to note that the seismicity is concentrated on the part of massif with-

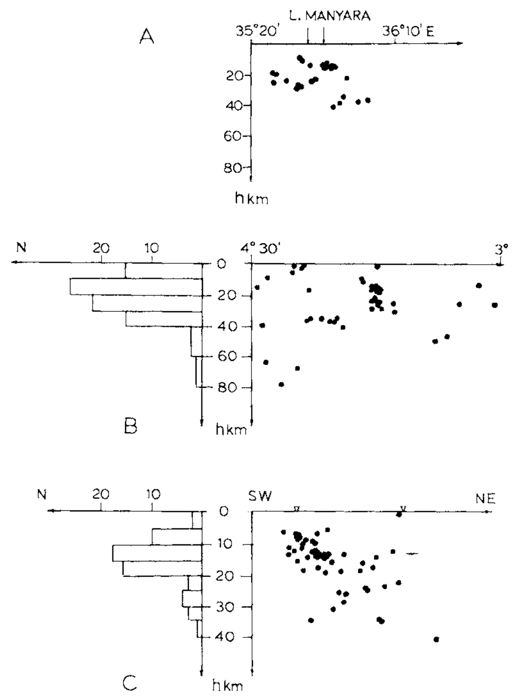

Fig. 2. Vertical distribution of earthquake foci. A. From east to west through Lake Manyara. B. Along the southern part of Gregory Rift. C. From the northern part of Ruwenzori to the southern part of Lake Albert.

out any visible geological features. The epicentres form a wide band across the rift valley.

Fig. 2C shows the vertical cross section along the rift from the northern end of Ruwenzori to the southern part of Lake Albert. It is easy to note the clear tendency for the depths of foci of the earthquakes to increase in this direction.

(2) The energy of the recorded earthquakes was classified using the method and nomograms which have been applied to the seismic regions of Pamir (Riznichenko, 1960). The energy of the earthquakes (E) may be expressed (Gutenberg and Richter, 1956) as:

$$K = \log E \text{ (joules)} = 4 + 1.8\,M$$

The energy range of recorded earthquakes was:

$$K = 5 - 10 \text{ or } M = 0.5 - 3.5.$$

Fig. 3 shows the level of seismicity and relation between number and energy of events for the regions under investigation (N is number of events per 100 km^2 per year, the areas

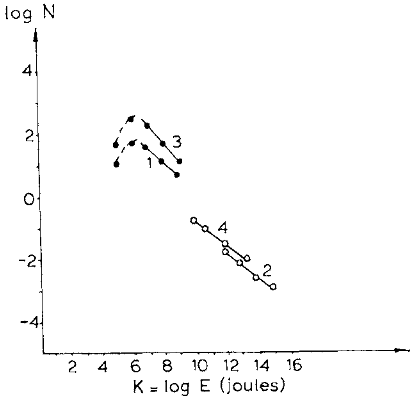

Fig. 3. Energy $K = \log E$ (joules) as a function of the number of earthquakes (N). *1*: for microearthquakes of the southern part of the Gregory Rift: *2*: for earthquakes with $K > 12$ (by Wohlenberg) of the same area (Wohlenberg, 1967); *3*: for microearthquakes of Ruwenzori; *4*: for earthquakes with $K > 10$ (by Wohlenberg) of the same area (Wohlenberg, 1967).

of analyses being indicated in Fig. 1B, C). 230 records of the earthquakes (10 days of observation) for Ruwenzori region (curve *3*) and 90 records (5 days of observation) for the southern part of Gregory Rift (curve *1*) have been included in the analysis.

(3) Some information about the deep structure of the rift zone has been obtained for the southern part of Gregory Rift. Seismograms of the earthquakes for this region sometimes had clear arrivals of different wave phases for which interpretation was possible (Fig. 4A). The seismograms of Ruwenzori region were more complicated.

Observational dependences:

$$(t_{P(REFL)} - t_{\bar{P}}) = f_1\,(t_{\bar{S}} - t_{\bar{P}});\ (t_{\bar{P}} - t_P) = f_2\,(t_{\bar{S}} - t_{\bar{P}})$$

where $t_{\bar{P}}$, $t_{\bar{S}}$, $t_{P(REFL)}$, t_P, are the times of arrival of straight P and S waves, and reflected and refracted P waves respectively (see Fig. 4B).

Hypocentral distance (D) can be expressed as:

$$D = K\,(t_{\bar{S}} - t_{\bar{P}})$$

where K is a function of velocities of P (V_P) and S (V_S) waves:

$$K = \frac{V_P \cdot V_S}{V_P - V_S} \quad \text{km/sec}$$

The value of k determined for depths of foci of about 20 km was (7.8 ± 0.4) km/sec in the

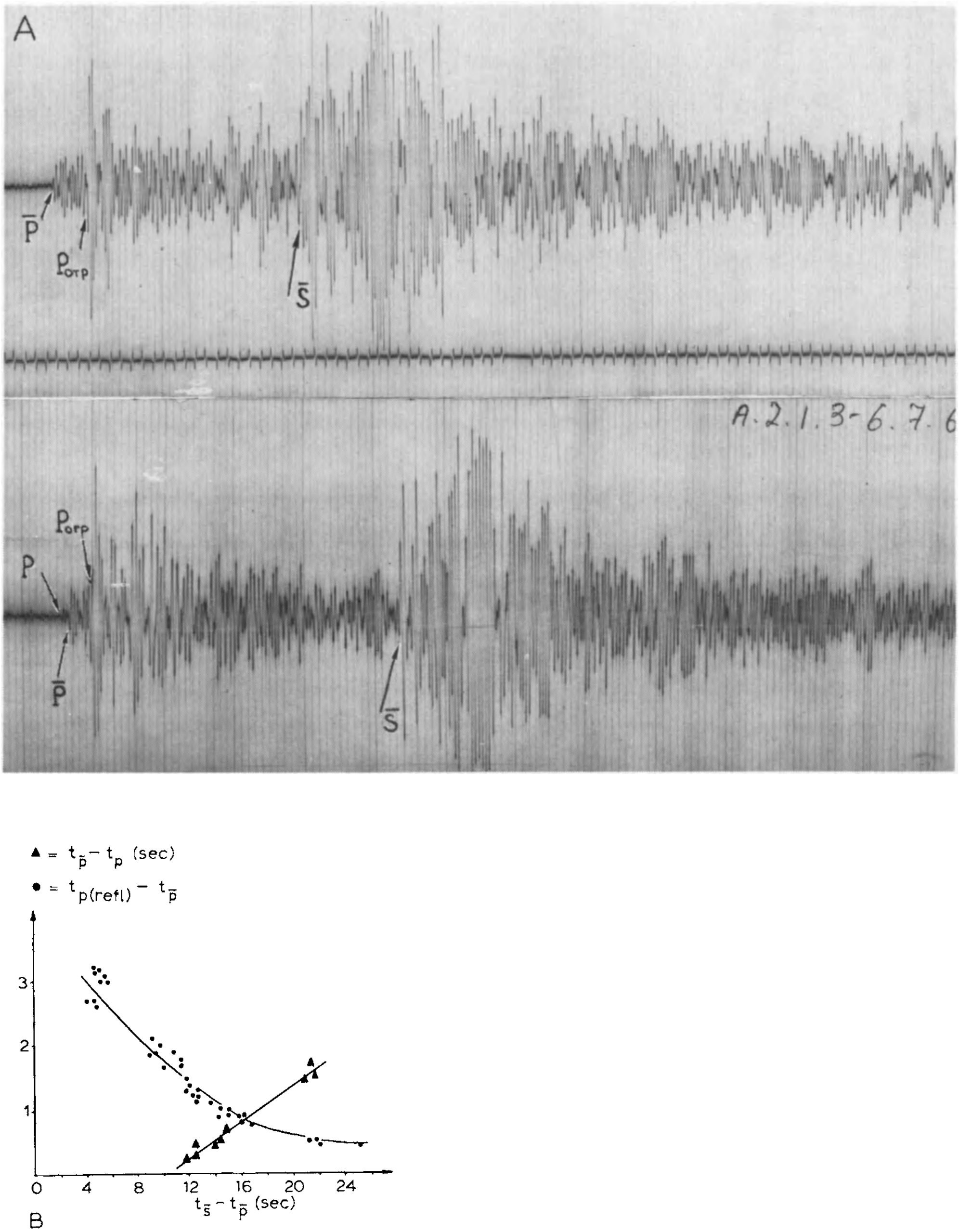

Fig. 4. A. Records of the earthquakes. B. Dependences $(t_{P(REFL)} - t_{\overline{P}}) = f_1\,(t_{\overline{S}} - t_{\overline{P}})$ and $(t_{\overline{P}} - t_P) = f_2\,(t_{\overline{S}} - t_{\overline{P}})$.

region under investigation.

Suppose that $V_P/V_S = 1.75$. Then:

$$V_{P(1)} = K\left(\frac{V_P}{V_S} - 1\right); \qquad V_{P(1)} = (5.8 \pm 0.3)\ \text{km/sec}$$

where $V_{P(1)}$ is the velocity of P waves in the upper 20 km of the crust. If the records of the earthquakes showed the arrivals of refracted P waves, when $t_{\bar{S}} - t_{\bar{P}} > 10$ sec (Fig. 4), these waves were connected with the intermediate discontinuity.

In our case:

$$\frac{d\,(t_{\bar{P}} - t_P)}{d\,(t_{\bar{S}} - t_{\bar{P}})} = \alpha \simeq K\left(\frac{1}{V_{P(1)}} - \frac{1}{V_{P(2)}}\right)$$

where $V_{P(2)}$ is the velocity of P waves below the intermediate discontinuity. The value of α is determined by observations ($\alpha = 0.135$ when $t_{\bar{S}} - t_{\bar{P}} > 10$ sec (Fig. 4)). Then:

$$V_{P(2)} = \frac{V_{P(1)} \cdot K}{K - \alpha V_{P(1)}} = (6.5 \pm 0.3)\ \text{km/sec}$$

The depth to the intermediate discontinuity can be obtained from the formula:

$$H_1 = \frac{[\alpha(t_{\bar{S}} - t_{\bar{P}}) - (t_{\bar{P}} - t_P)]\ V_{P(1)} \cdot V_{P(2)}}{2\sqrt{V^2_{P(2)} - V^2_{P(1)}}} + \frac{h}{2}$$

where h is the depth of focus of the earthquake. If we take into account the distribution of depths of foci in the area (Fig. 2B) it will be possible to say that:

$$H_1 = 18\ \text{km}$$

Some information about the thickness of the earth's crust can then be obtained by analysis of dependence:

$$(t_{P(REFL)} - t_{\bar{P}}) = f_1\ (t_{\bar{S}} - t_{\bar{P}})$$

Using theoretical curves for different models of the earth's crust, the value:

$$H_2 = 35 \text{ to } 37\ \text{km}$$

is found to be in best accordance with the observational data.

Therefore the earth's crust at the southern end of the Gregory Rift is near to standard structure (thickness of upper layer is $\sim$ 18 km, total thickness of the crust is $\sim$ 36 km; velocities of P waves are (5.8 ± 0.3) km/sec and (6.5 ± 0.3) km/sec in the upper and intermediate layers respectively.

CONCLUSIONS

The principal aims of this work were to estimate numerically some interesting parameters of rift zones as a basis for larger scale investigations.

(1) A close correlation between the distribution of epicentres of small earthquakes and geological features of rifts was not found *.

(2) The seismicity of the rifts (in range of microearthquakes) shows that it is high. It is possible to compare it with the seismicity of other very active seismic zones of the earth (with Pamir for example). It is interesting to note that values of magnitudes of strong earthquakes in areas under investigation are limited by $M = 7.25$ (Drake, 1969).

(3) It has been found that the structure of the earth's crust in the southern part of the Gregory Rift is near to typical for continents.

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The authors are very grateful to Prof. V.V. Beloussov and Prof. A.P. Kapitza – leaders of Soviet East African Expedition.

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* All foci of the earthquakes were situated inside the crust (the depths of foci were less than 35–40 km).

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GRAVITY AND MAGNETIC SURVEYS IN NORTHERN TANZANIA AND SOUTHERN KENYA

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ABSTRACT

Darracott, B.W., Fairhead, J.D. and Girdler, R.W., 1972. Gravity and magnetic surveys in northern Tanzania and southern Kenya. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 131–141.

The main results of gravity surveys carried out in 1968 and 1969 are presented with occasional reference to magnetic surveys carried out at the same time. The long wavelength negative Bouguer anomaly associated with the northern part of the rift zone is found to terminate at about 4°S and the shorter wavelength positive anomaly over the Gregory Rift floor dies out at about 2°S. The negative anomaly is thought to be due to a low density asthenolith and the positive anomaly to be an intrusive zone penetrating the upper crust suggesting extreme thinning of the lithosphere beneath the eastern rift north of 2°S. The negative Bouguer anomaly over the Speke Gulf region is interpreted in terms of a simple rift; reasons are given for considering this to be a Precambrian structure which is presently being rejuvenated. The negative Bouguer anomalies found over the west-east volcanic chain which includes Meru and Kilimanjaro are considered to be due to low density lava piles.

INTRODUCTION

The gravity and magnetic surveys reported here cover parts of the regions of rifting and volcanism in northern Tanzania and southern Kenya (Fig.1), i.e., the southern limit of the belt of continuous volcanicity associated with the Gregory Rift. The rift system started to form in Miocene times and there has been a complex sequence of updoming, volcanism, and faulting (see, e.g., Gregory, 1921; UMC/UNESCO, 1965). In southern Kenya, the volcanic activity is mainly confined to the rift trough. The Precambrian basement is exposed along the rift scarps. The Precambrian crust can be divided into: (1) the Tanganyika shield, to the west of longitude 35°E, comprising a large area of migmatites and mobilized granites; and (2) the Mozambique orogenic belt to the east, composed of metamorphosed sedimentary rocks (Cahen and Snelling, 1966). Tertiary faulting is mainly confined to the rift trough but also occurs at a few places away from the rift. In northern Tanzania, the faulting changes from a simple rift to a series of block faults with east facing scarps. Only the western wall of the Kenya rift continues into Tanzania as a major feature. The history of the volcanic province in northern Tanzania, based on potassium-argon age determinations, has been summarised by Evans et al. (1971).

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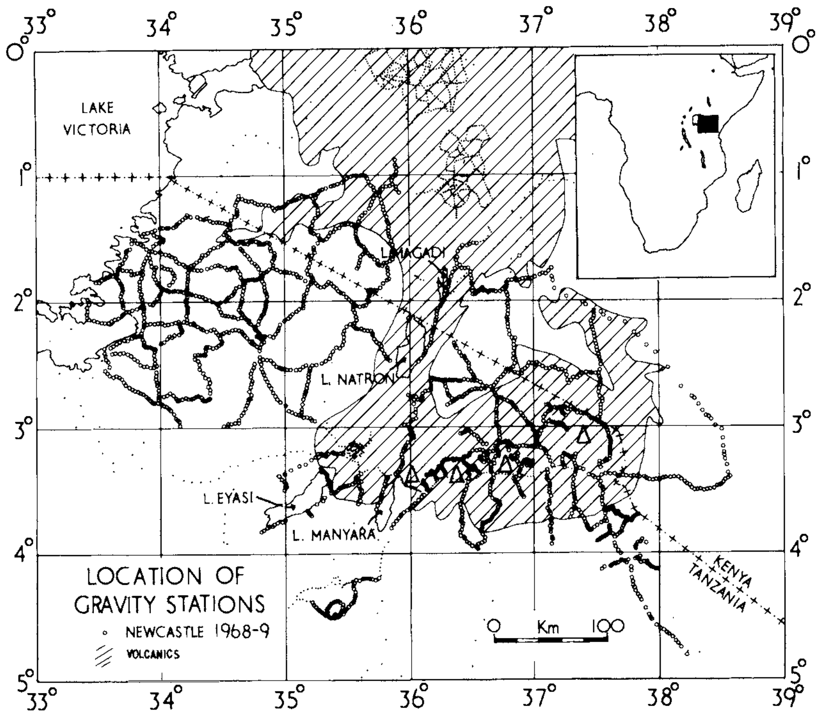

Fig.1. Location of gravity measurements in northern Tanzania and southern Kenya. Other sources of data include Masson Smith and Andrew (1960), Searle (1970) and J. Mansfield (personal communication, 1970). The shaded area represents the volcanic province, and the open triangles the volcanic centres.

The objects of the survey were: (1) to investigate the nature of the broad negative Bouguer anomaly associated with the Eastern Rift (Bullard, 1936; Girdler et al., 1969; Girdler and Sowerbutts, 1970); (2) to investigate the southern termination of the axial positive anomaly in the Kenya Rift, previously mapped further north by Searle (1970); (3) to investigate the relationship of the Speke Gulf to the Gregory Rift; and (4) to investigate the relationship between the rift system and the west–east line of volcanoes in northern Tanzania which includes Kilimanjaro and Meru.

THE GRAVITY SURVEY

2243 measurements were made along roads and tracks in 1968 and 1969 (Fig.1). The measurements were made using a Lacoste–Romberg geodetic gravity meter and were tied to the base station network of Masson Smith and Andrew (1962). Elevations were obtained using a "leap-frog" barometric levelling technique (Searle, 1969), and were tied into trigonometric points and bench marks where possible. Free-air and Bouguer anomalies (for S.G. = 2.67) were calculated. Terrain corrections were computed to a distance of 166.7 km for stations east of the Natron–Manyara escarpment (longitude 36°E). West of

this, the topography is relatively smooth and the terrain corrections were neglected as they were considered to be less than 10 g.u.

INTERPRETATION OF THE GRAVITY ANOMALIES

The free-air anomalies fluctuate about zero, with wavelength similar to the change in topographic relief, indicating that the region is in approximate isostatic equilibrium.

The Bouguer anomaly (Fig.2) decreases from –1400 g.u. along the shores of Lake Victoria in the west to –1900 g.u. over Lakes Eyasi and Magadi, and then increases again to –500 g.u. in the east. In the north of the survey area, the Bouguer anomaly reaches its

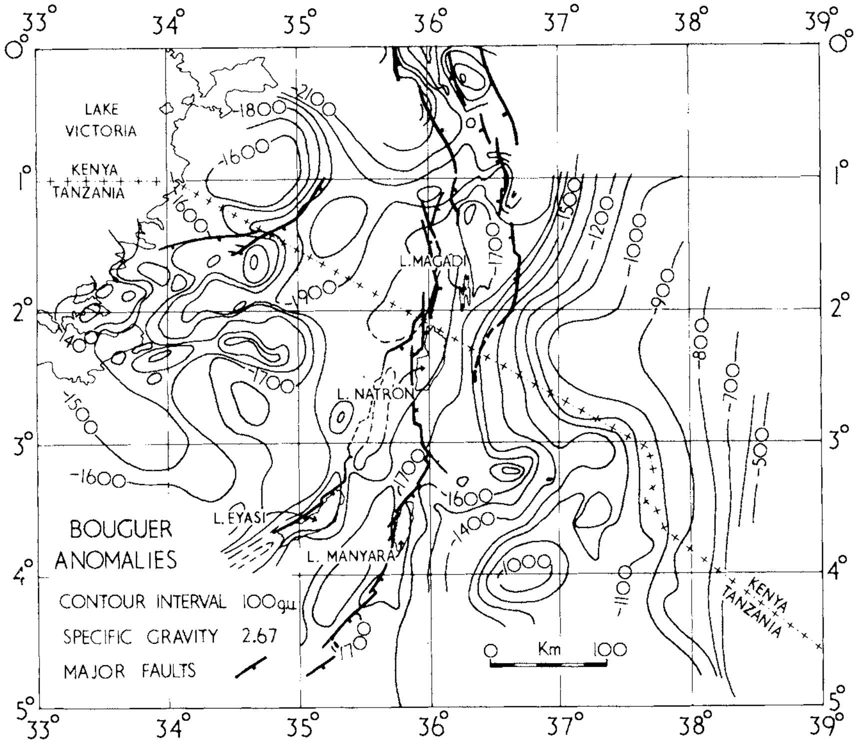

Fig.2. Contour map of the Bouguer anomalies (equatorial Mercator projection).

most negative value of –2400 g.u., and correlates with the area of maximum uplift of the Kenya dome (Baker and Wohlenberg, 1971), while further south, the local negative anomalies parallel the downthrow side of the major rift faults.

The regional Bouguer anomaly

The Gregory Rift is associated with a broad negative Bouguer anomaly, of about 500 km width and amplitude –1000 g.u. This has been interpreted by Girdler et al. (1969) as

being due to the slightly lower density asthenosphere expanding to higher levels and engulfing part of the lithosphere. The lithosphere (uppermost mantle plus crust) is thus thinned under East Africa, a situation somewhat similar to that found in the Basin and Range province of the western United States. Girdler and Sowerbutts (1970) produced a Bouguer gravity map for East Africa by averaging all the available data by 2 X 2 degree squares. They suggested this might give some idea of the extent of the region of thinning. The long wavelength Bouguer anomaly appeared to die out south of about 5°S, near where the Gregory Rift changes from a complex graben with volcanism to a more simple region of block faulting. The present survey allows the termination of the negative regional Bouguer anomaly to be delineated more accurately. The regional Bouguer anomaly shown in Fig.3 was obtained by smoothing ten long profiles, and contouring. The anomaly is seen to die out at about 4°S. The overall trend of the anomaly is N25°, rather than the north–south trend of the rift valley.

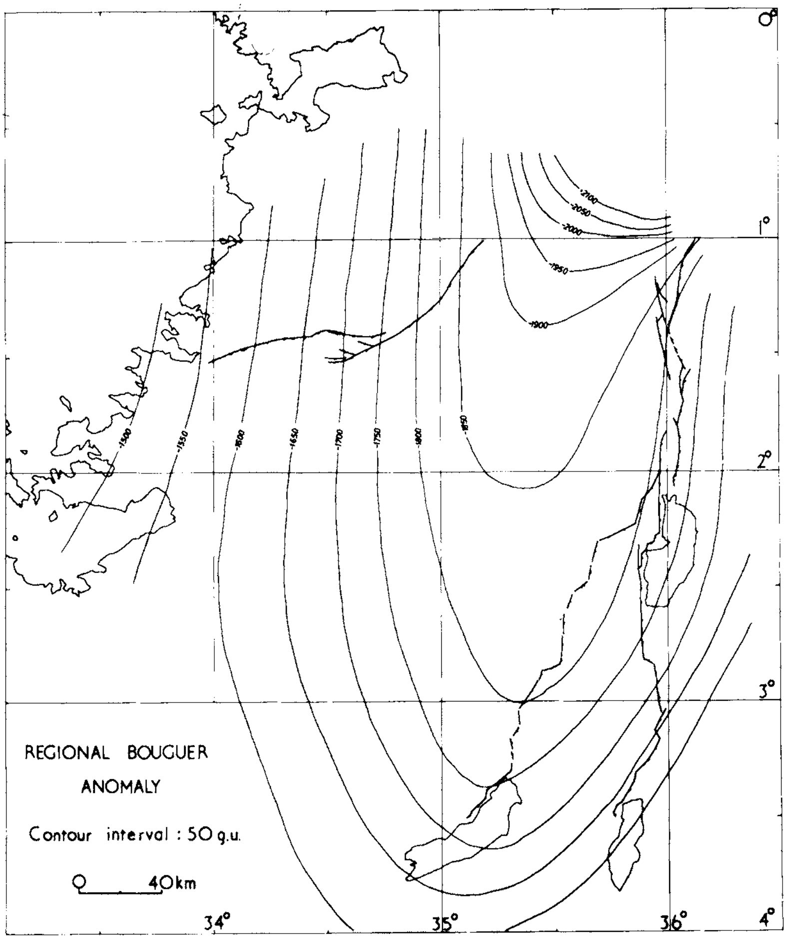

Fig.3. Contour map of the regional Bouguer anomaly (equatorial Mercator projection). The specific gravity used is 2.67.

Fig.4 shows a quantitative interpretation of the regional Bouguer anomaly for a west–east profile at the latitude of Magadi (1.8°S). With the aid of a computer programme of Takin and Talwani (1966) several models have been tried and the one shown assumes the lithosphere to be about 90 km thick including a crust of 35 km. The regional crustal structure assumed is based on that derived by Gumper and Pomeroy (1970). Their model,

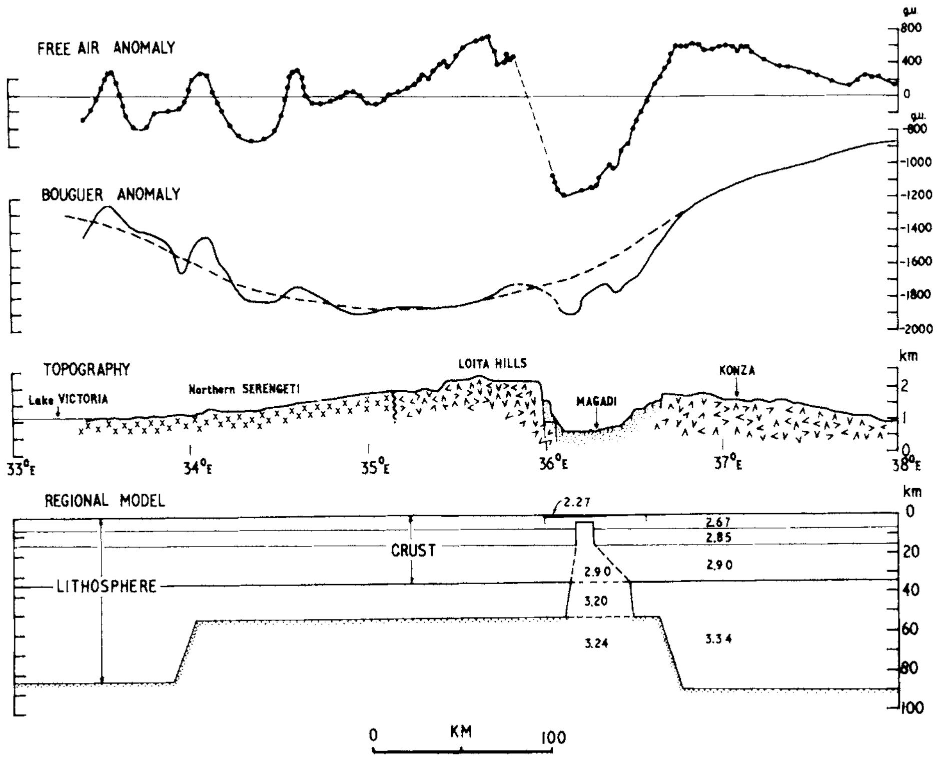

Fig.4. Gravity anomalies, geology and model of the lithosphere and asthenosphere satisfying the gravity for a section in the latitude of Magadi (1.8°S). The dashed line in the Bouguer profile is the assumed regional.

from seismic studies consists of a three-layer crust with the top 7 km having S.G. = 2.70, the next 10.5 km having S.G. = 2.80, and the lowest 18.7 km having S.G. = 2.85. These values have been slightly altered by giving a specific gravity of 2.67 to the upper layer and keeping a mean crustal specific gravity of 2.84, to conform with the gravity reduction procedure. The background Bouguer anomaly over the normal lithosphere is assumed to be –900 g.u. following Girdler et al. (1969). The Kenya coast is over 500 km to the east of the Gregory Rift, at this latitude, and its effects are considered small. The negative anomaly is again assumed to be due to a low density asthenolith, the specific gravity of which is envisaged to vary both vertically and horizontally. For ease of computation, a mean specific gravity contrast of –0.10 is assumed. With these assumptions, the asthenolith is found to be about 290 km wide and to reach to within about 50 km of the surface (Fig.4).

The axial positive anomaly in the Gregory Rift

Removal of the regional Bouguer anomaly (dashed line in Fig.4) over the Magadi area leaves a negative residual Bouguer anomaly with a smaller central positive anomaly superimposed. The negative residual anomaly over the rift floor correlates with the distribution of the low density lavas within the rift, and the central positive anomaly is similar to the anomaly found further north in Kenya (Searle, 1970) and Ethiopia (Gouin and Mohr, 1964). The present work shows that the axial positive anomaly extends as far south as, and ends near the eastern shore of Lake Magadi. To explain the anomalies, we adopt a similar model to that proposed by Searle (1970). Corrections have been made for the lavas within the rift which have a specific gravity contrast of –0.4, requiring their thickness to be approximately 1.5 km. Beneath these, and intruding the upper crust, is the denser "gabbroic" body with a specific gravity of 2.9. The intrusion comes to within 4 km of the surface and has a width of 10 km. This model should be compared with the work further north, where Searle (1970) was able to show that the dense intrusive zone beneath the axis of the rift is about 20.km wide, reaching in places to within 3 km of the rift floor. This type of model accounts for the correlation between the axis of the positive anomaly and the geothermal activity, grid faulting and volcanism (Searle, 1970).

The model in Fig.4 is also in accord with the results of Fairhead and Girdler (1971), who, in a study of the seismicity of Africa and station travel time corrections, noticed that the Eastern Rift north of about 4°S is associated with a slowing down of P waves; they suggested that the regions of P delay, S_n attenuation, and the long wavelength negative Bouguer anomaly are all related.

The gravity anomalies over the Speke Gulf region

To the west of the Gregory Rift (Fig.5) the Precambrian basement is a complex granitic craton, containing the relics of the Nyanzian orogeny (c. 3000 m.y.). This complexity is reflected in the complexity of the Bouguer anomalies. In order to interpret the anomalies, the effects of the near-surface crustal features (local anomalies) have been separated from those arising from deeper causes (regional anomalies). The resulting residual Bouguer anomalies are shown in Fig.6, in which the north–south trending dashed line (after Cahen and Snelling, 1966) marks the boundary between the Tanganyika shield to the west, and the Mozambique orogenic belt to the east. Over the Mozambique belt there are no large residual Bouguer anomalies, and in general the gravity field is smooth. In contrast, over the shield there are several pronounced residual Bouguer anomalies. The same observation holds for the magnetic anomalies. This indicates that the residual anomalies over the shield are probably caused by structures which are at least as old as the Mozambique orogeny (450–600 m.y.), and that if similar structures ever existed further to the east, then they have been reworked and largely obliterated by the Mozambique orogeny. The coincidence of the change from a craton to a mobilised orogenic belt and the change in character of the residual Bouguer anomalies is particularly impressive, and it seems that gravity and magnetic surveys can be usefully employed to delineate them.

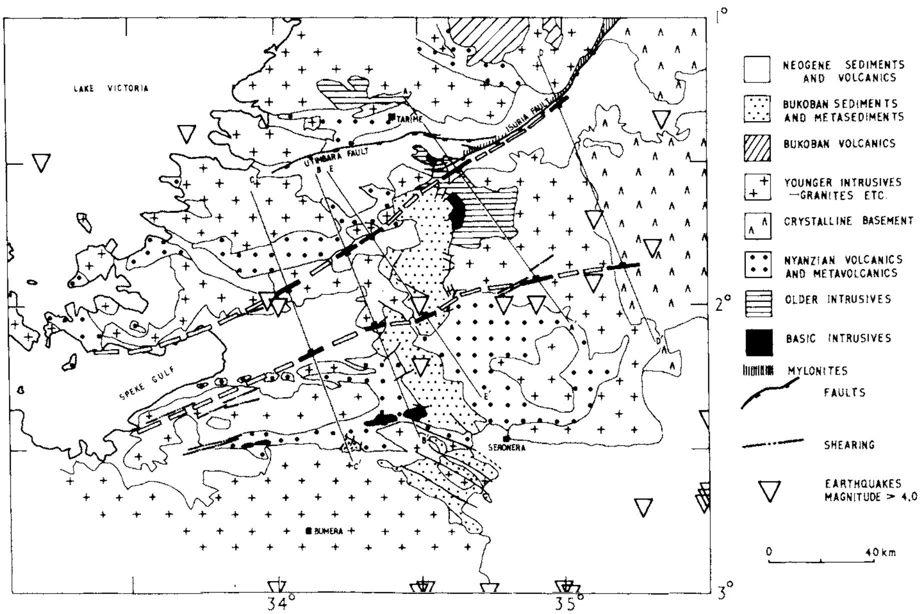

Fig.5. The position of the faults (broad dashed lines) inferred from the five profiles, superimposed on the geologic map of the Speke Gulf region, showing the correlation with shear zones and the seismicity.

Over the Tanganyika shield, the positive residual Bouguer anomalies correlate well with the surface geology; for example, the near-circular positive anomaly centred at 1.6°S, 34.7°E correlates with the outcrop of basic intrusives and 60 km to the south, the large positive anomaly is related to the relatively high density greenschists of the Nyanzian System.

The prime object of this part of the survey is to seek some relationship (if any) of the Speke Gulf to the main Gregory Rift. It is seen that a trough-shaped negative residual Bouguer anomaly (–200 g.u.) aligns with the Speke Gulf (Fig.6). The length of the anomaly is about 200 km and its width varies from 45 to 60 km. After the positive residual anomalies have been separated, the negative residual anomaly is found to be not well correlated with the *surface* geology. We consider three possible causes for this negative residual anomaly:

(1) *Low density Tertiary and Quarternary sedimentary cover*

The sediments near the Gulf do not follow the axis of the negative anomaly (Fig.5) and in any case are believed to be less than 100 m thick. Such a thickness would give a maximum gravitational affect of only –35 g.u. (c.f. the observed –200 g.u.) for an assumed specific gravity of 1.8 for the sediments.

(2) *A large granitic intrusion*

The granites within the area of the anomaly are of similar density and mineralogy to those elsewhere, and a low density intrusion at depth would need to have a specific gravity

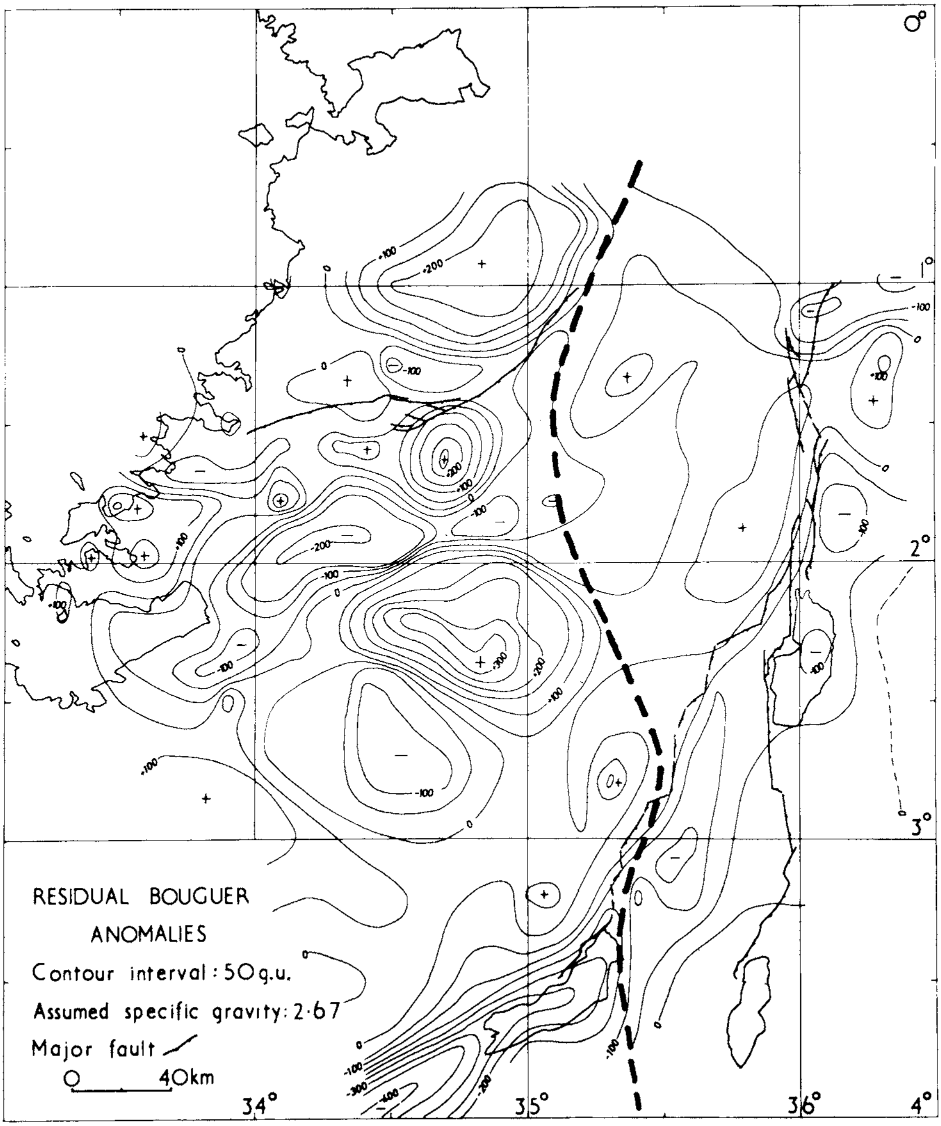

Fig.6. Residual Bouguer anomalies over the Speke Gulf region. The north–south dashed line represents the boundary between the Tanganyika shield to the west and the Mozambique orogenic belt to the east.

as low as 2.57 to produce the observed anomaly. This is very low for a granite. A granite intrusion with a linear shape measuring 200 km by 50 km also seems very unlikely.

(3) *Speke Gulf is a graben*

The Gulf is somewhat similar to the Kavirondo Gulf in Kenya, and Horne (1962a, b) has suggested that the Speke Gulf may be a Precambrian graben structure. This explanation

is favoured here. The negative anomaly crosses the north–south belt of Bukoban sediments (c. 1000 m.y.) and these are not appreciably disturbed by faulting. The proposed rift must, therefore, be older than this formation. The granites in the area have been dated at c. 2500 m.y. (Cahen and Snelling, 1966; Edwards and Howkins, 1966) and the rift must obviously be younger than this. The region is very eroded and the location of any faults is unclear. It is assumed that the rift developed in a layered crust in the Precambrian and the fault scarps eroded away, leaving a relatively uniform surface. This results in lower density rocks within the rift in juxtaposition with higher density material at the sides, thus producing the required negative density contrast. Fig.7 shows the model for the NW–SE

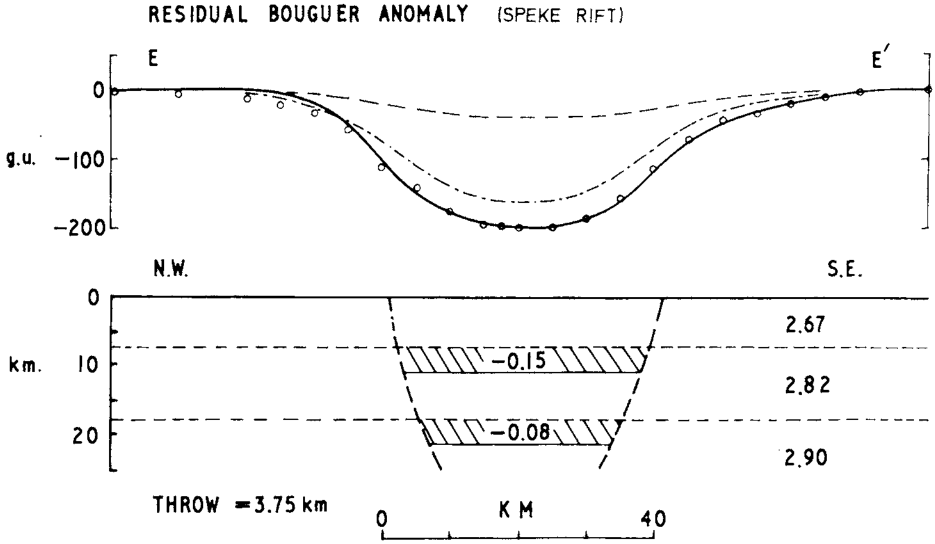

Fig.7. Model for the Speke Gulf Rift (?Precambrian), for profile E–E′. The regional crustal structure is slightly modified after Gumper and Pomeroy (1970). The heavy black line is the observed anomaly; the chained line is the anomaly due to the upper density contrast; the dashed line is the anomaly due to the lower density contrast; and the black dots the total computed anomaly.

profile E–E′ (Fig.5). A similar analysis on the other four profiles across the negative anomaly allows the approximate position of the inferred faults to be mapped (Fig.5). The postulated faults correlate well with shear zones in the granitic rocks, for example, the zone of the Precambrian mylonites along the Isuria escarpment may be an expression of a former fault line. The maximum vertical displacement computed for the rift is about 4 km. This seems reasonable as Precambrian rifts in other shield areas are known which have vertical displacements of this order (e.g., Kanasewich et al., 1969).

It is proposed that the inferred Precambrian rift has acted as a preferential site for the position of the Miocene Isuria fault, and the possible Tertiary features of the Speke Gulf. Wohlenberg (1968, 1970) found a northeast trending linear zone of earthquake epicentres ($M \geqslant 4$) 50 km to the south of the Isuria fault (Fig.5). It is further proposed that the

seismicity indicates that the southern inferred fault is being rejuvenated. The faults are almost certainly normal, and may have small components of strike-slip motion, that could be caused by differential movements between regions to the north and south of the zone, in response to the north-south variation in the opening of the Gregory Rift.

The gravity anomalies associated with the volcanic province of northern Tanzania

To the east of the Natron–Manyara escarpment, the volcanic province is dominated by the east–west line of volcanic centres (Fig.1). The trend of the negative Bouguer anomaly associated with the volcanic centres is clear from Fig.2. Density profiling (Nettleton, 1939) along north–south traverses over the volcanic chain, and density measurements suggest the negative anomaly is caused by the low density lava piles associated with the volcanic centres. Similar conclusions were reached by Searle (1969) for the Ngorongoro Caldera in Tanzania, and by Khan and Mansfield (1971) for the Mount Elgon and Mount Kenya volcanoes of Kenya.

The east–west volcanic chain has been described as a possible "leaky transform fault" (Evans et al., 1971). This could possibly take up some of the north–south variations in the extension across the Gregory Rift.

CONCLUSIONS

This survey has shown that the region of lithospheric thinning beneath East Africa dies out at about 4°S, and that the zone of severe thinning along the axis of the Gregory Rift, manifested by the axial positive residual Bouguer anomaly, ends at about 2°S.

The Speke Gulf is interpreted as a Precambrian rift, and the seismicity suggests it is being rejuvenated. This seismicity, and the volcanoes of northern Tanzania may possibly result from north–south variations in the extension across the Gregory Rift.

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REPORT ON AIRMAGNETIC SURVEYS OF TWO AREAS IN THE KENYA RIFT VALLEY

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(Received February 1, 1972)

ABSTRACT

Wohlenberg, J. and Bhatt, N.V., 1972. Report on airmagnetic surveys of two areas in the Kenya Rift Valley. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 143–149.

Airmagnetic surveys have been carried out in two areas within the Kenya Rift system: the Lake Magadi region and the region south of Lake Hannington. In both regions, but particularly in the area south of Lake Hannington, NW–SE trending magnetic anomalies stand out clearly. The source of these anomalies is considered to be in the deeper earth's crust beneath the rift zone. At present no explanation for the anomalies can be offered. There is no convincing evidence from the magnetic data for the occurrence of extended intrusive bodies in the upper part of the earth's crust beneath the rift floor.

INTRODUCTION

The Kenya Rift Valley is part of the major East African continental rift system that extends from the Afar triangle to North Tanzania. Baker and Wohlenberg (1971) showed that rift formation has not been the result of crustal updoming and faulting alone, but that both doming and faulting must be related to some subcrustal process involving igneous intrusion.

Interpretation of gravity data for the Kenya Rift suggests the existence of large basic intrusions in its central and southern parts (Sowerbutts, 1969; Searle, 1970; Baker and Wohlenberg, 1971; Darracott et al., 1972). Support for the presence of large intrusions beneath the Kenya Rift also comes from seismic investigations (Wohlenberg, 1969, 1970; Griffiths et al., 1971).

The purpose of the airmagnetic investigations described in this report was to find out whether it is possible to detect by magnetic methods the top of such an intrusive body beneath the extensive lava flows which cover most parts of the rift floor.

PLANNING OF THE AIRMAGNETIC SURVEY

No previous airmagnetic investigations within the Kenya rift system have been carried out. Therefore, special care was taken in choosing the regions and the method of the

survey, to ensure best results.

Two regions were chosen for the survey which fulfilled the conditions that detailed geological reports and maps were available, and the structure of the rift floor of these two regions is not masked by recent deposits.

The two regions selected were the Lake Magadi Area (*AM*) and the Lake Hannington Area (*AH*) (Fig.1).

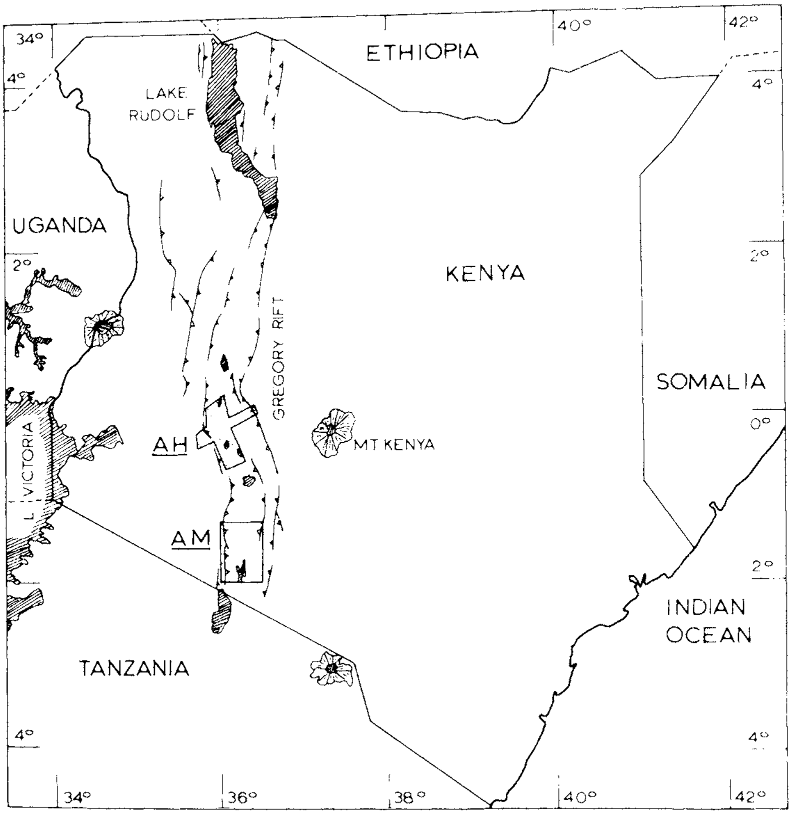


Fig.1. Location of the surveyed areas. *AM* = Airmagnetic Magadi; *AH* = Airmagnetic Hannington.

To find the ideal conditions for the flight altitude and flight direction, models were calculated using computers. These calculations were based on the assumption that basic materials had intruded the earth's crust and that these bodies are dyke shaped and that their strike directions are parallel to the main tectonic lines of the rift, i.e., N–S, and are situated within it. For the external inducing magnetic field, data of the earth's actual magnetic field were taken regarding the local magnetic coordinates. Information on the remanent magnetisation was obtained from studies of orientated rock samples (A. Brock, personal communication, 1970).

As a result of these calculations it became apparent that the best flight direction

would be NE–SW. The spacing of the profiles was chosen to be 5 km.

In view of different topographic relief in each region it was decided to carry out the survey of the Lake Magadi Area at a constant flight altitude above sea level (A.S.L.), while the Lake Hannington area should be flown at a constant ground clearance (A.G.L.). The Magadi area on average shows a much smoother topography than the Lake Hannington area which is characterized by steep N–S trending fault scarps.

The flying programme was carried out on October 15, 1970 (Lake Magadi area) and on October 16, 1970 (Lake Hannington area). A total of 1870 line-kilometres were flown, 865 km in the Lake Magadi area and 1005 km in the Lake Hannington region. The magnetic data have been corrected for the diurnal variations. No reductions have been made which take into account the regional field and the topography. First calculations show that neither the regional field nor the topography will affect the results appreciably.

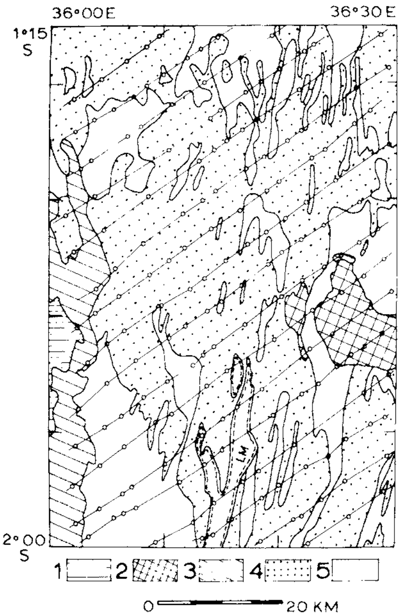
AIRMAGNETIC SURVEY OF THE LAKE MAGADI AREA

The area which is covered by the first part of the survey and which is named after the soda Lake Magadi is situated between 1°15′S, 2°00′S, 36°00′E and 36°30′E. The total area covered was approximately 3750 km^2. The survey was flown at a constant flight altitude of 2000 m (ca. 6000 ft) above sea level which results in an average flying level above ground of approximately 1000 m (3000 ft). The direction of the profiles is NE–SW, its total number 17 (Fig.2 and 3).

The geology of the area south of 1°30′S was mapped by Baker (1958). The region north of 1°30′S is covered by a geological map compiled by Randel (1970). The exposed rocks in the area (Fig.2) consist of Pliocene flood basalts resting on Precambrian basement in the west and a central volcano to the east. Most of the rift floor of this area is covered by Pleistocene to Quaternary trachytes and lacustrine sediments. All these formations are cut by N–S trending faults of various ages.

The magnetic data obtained are presented as a set of profiles (Fig.3). A base value of the total field $T = 34{,}550\ \gamma$ has been chosen. In the presentation the flight profiles are taken at the same time as base lines. The average anomalies in T vary from $+200\ \gamma$ to $-200\ \gamma$ relative to the base value. Extreme values are found along profile Q with a positive anomaly of $+400\ \gamma$ and along profile F with a negative anomaly of $-800\ \gamma$.

Besides N–S striking magnetic alignments which are more numerous in the northern part of the area there are some dominant NW–SE trending anomalies. A half wave length of about 20 km of these latter anomalies suggest their origin to be in the deeper earth's crust. While the strong negative anomaly in the center right of the area is related to a volcano (Olorgesailie) no evidence has been found until now of any other correlation between surface features and observed magnetic anomalies. There is no convincing evidence for extended N–S trending intrusions.



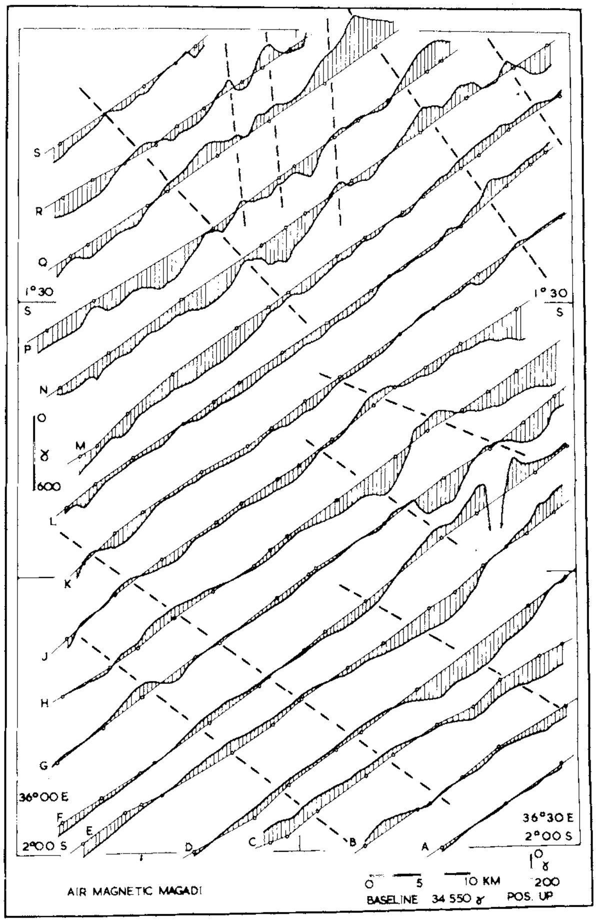


Fig.2. Geology of the Lake Magadi Area after Baker (1958) and Randel (1970). *1* = Precambrian basement; *2* = Pliocene central volcano; *3* = Pliocene basalts; *4* = Plio–Pleistocene trachytes; *5* = Quaternary sediments and pyroclastics; *LM* = Lake Magadi. Flightlines plotted. Open circles are control points along the profiles.

Fig.3. Set of magnetic records taken along the 17 profiles of the airmagnetic survey of the Lake Magadi area. Base value $T = 34{,}550\ \gamma$. Anomalies relative to the base value are plotted below and above – not perpendicular to – the corresponding location on the profile lines which are taken at the same time as base-lines.

AIRMAGNETIC SURVEY OF THE LAKE HANNINGTON AREA

The area which is referred to as *AH* in Fig.1 stretches from Lake Hannington to south of Lake Nakuru and is bounded by 0°15′N, 0°30′S, 35°45′E and 36°30′E. A total of 1005 km was flown along 18 profiles (Fig.4, 5) keeping a constant ground clearance of ca. 600 m.

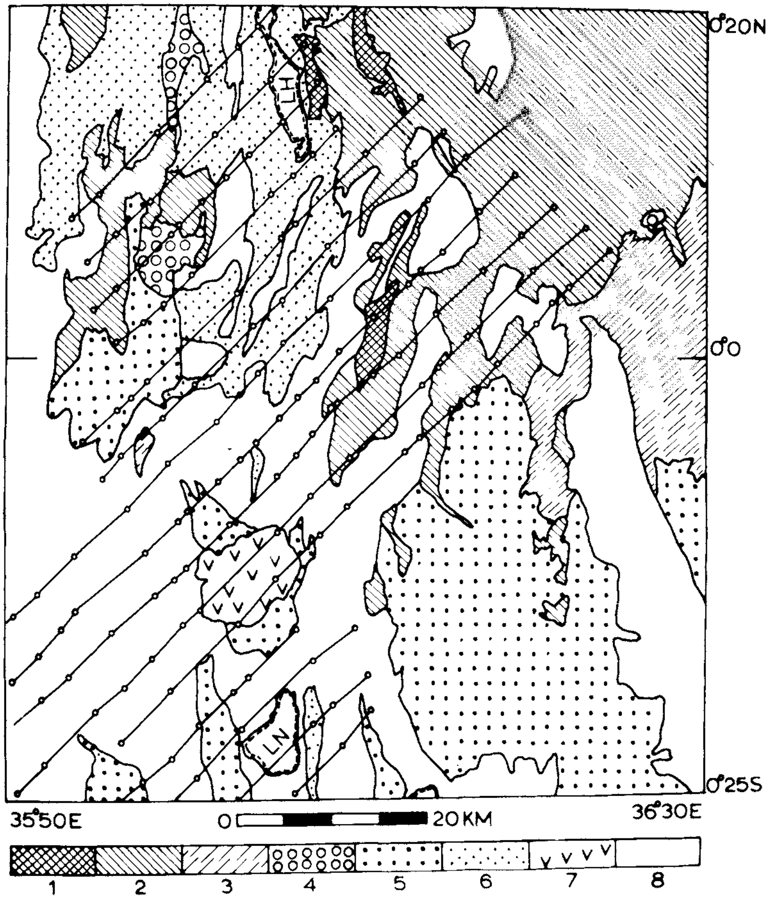


Fig.4. Geology of the Lake Hannington Area after Thompson and Dodson (1963), Jennings (1966), McCall (1967) and Walsh (1969).
1 = Miocene basalts; *2* = Upper Miocene phonolite; *3* = Mid-Pliocene phonolite; *4* = Pliocene basalts; *5* = Plio-Pleistocene trachytes and phonolites; *6* = Plio-Pleistocene ignimbrites; *7* = Quaternary volcanics; *8* = Quaternary sediments and pyroclastics. *LN* = Lake Nakuru; *LH* = Lake Hannington. Flight lines plotted. Open circles are control points along the profiles.

The geology of the area (Fig.4) is described by Thompson and Dodson (1963), Jennings (1966), McCall (1967) and Walsh (1969). The rocks exposed in the area are mainly Miocene to Quaternary alkaline volcanics, ranging from basic to phonolitic and trachytic lavas and tuffs. The region is extensively faulted with a few volcanic cones and craters. The major faults are approximately N–S.

The reduced magnetic data are again presented as a set of profiles (Fig.5) with the base value being $T = 34{,}500\ \gamma$. Even more striking than in the Magadi area is the appearance of dominant NW–SE trending alignment of magnetic anomalies. The half wave length of these anomalies again indicates a causative source in the deeper earth's crust. The magnitudes of the observed anomalies vary from $-100\ \gamma$ to $+200\ \gamma$. A strong magnetic low south of 0°15′S probably has to be correlated to a volcano (Menengai). Again in this surveyed area there is no clear evidence for extended N–S striking intrusive bodies. An explanation of the NW–SE trending magnetic anomalies remains open.

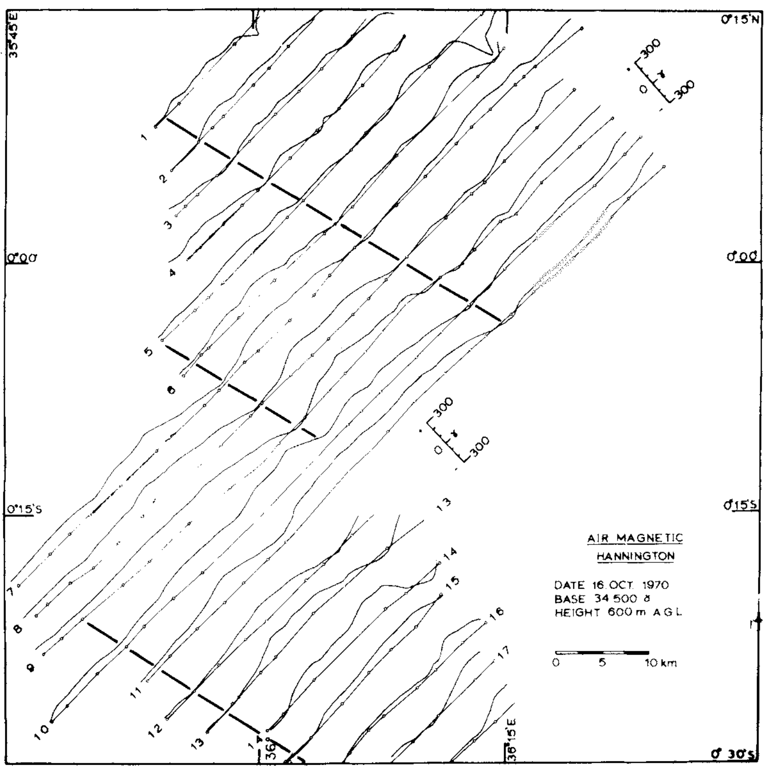


Fig.5. Set of magnetic records taken along the 18 profiles of the airmagnetic survey of the Lake Hannington area. Base value $T = 34{,}500\ \gamma$. Values plotted perpendicular to the flight lines.

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SOME COMMENTS ON THE RESULTS OF A SEISMIC REFRACTION EXPERIMENT IN THE KENYA RIFT

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(Received February 1, 1972)

ABSTRACT

Griffiths, D.H., 1972. Some comments on the results of a seismic refraction experiment in the Kenya Rift. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 151–156.

A reversed seismic refraction line was shot in the northern part of the Kenya Rift in 1968, Lake Hannington and Lake Rudolf being used for firing charges. Data quality was such that only a simple interpretation which assumed plane horizontal layering was carried out, though both first "P" and first "S" arrivals were used. The crust, with the high compressional wave velocity of 6.4 km/sec was found to overlie a layer with a velocity of 7.5 km/sec at a depth of about 20 km. However, arrivals from shots in Lake Hannington are not inconsistent with the presence of a basic intrusion along the centre of the rift, originally postulated to explain the presence of an axial gravity high.

STRUCTURE

The Gregory Rift of Kenya is a complex graben 60–70 km wide bisecting a large dome-shaped uplift. It is described by Baker and Wohlenberg (1971) as bounded by major normal faults arranged en echelon, with throws which may reach 3,000–4,000 m. In places a thickness of perhaps more than 2,000 m of volcanic lavas and sediments covers the rift floor, but to the north and south of the culmination of the Kenya dome, Precambrian rocks outcrop across virtually the whole of the rift zone, and the inference is that sialic crustal rocks are also present elsewhere beneath the volcanics.

Gravity measurements have indicated a long wavelength negative Bouguer gravity anomaly over the whole East African rift zone. This has been explained as due to thinning of the crust and lithosphere by separation of an East and West African plate (Girdler et al., 1969).

An axial high in the Gregory rift, noted by Searle (1970) and Khan and Mansfield (1971) has been interpreted both by Searle and by Baker and Wohlenberg as due to intrusion of basic material to shallow depth beneath the rift, believed to be associated with crustal thinning and perhaps a few kilometres of separation. The positive anomaly has an amplitude of some 50 mgal and has been traced from the region of the volcano Suswa in the south to Lokori in the north. It is thought that it may extend further, perhaps as far as Lake Rudolf, but to the east and some way south of Lodwar, another strongly positive area occurs. The

suggestion has been made by M.A. Khan that alternatively, south of the high, there may be a very considerable offset of the rift axis to the west across a fault trending at right angles to the axis. Since no major east-west structure displaces the rift valley at this point the fault would have to be of transform type, implying crustal separation of tens of kilometres in the rift floor.

SEISMIC REFRACTION EXPERIMENTS

The main purpose of the seismic refraction experiment carried out in the Gregory Rift (Griffiths et al., 1971) was to measure crustal thickness, in order to provide some control for the interpretation of the available gravity data, and to test the hypothesis of crustal thinning. As originally planned the measurements were to include long refraction lines both along the rift and across it, but for various practical reasons the transverse line was not fired.

The only lakes that could be used for shooting were Lake Hannington and Lake Rudolf, the refraction line thus being limited to the northern and somewhat inaccessible part of the Eastern Rift. The shot points and the station positions, sited at about 30 km apart on the only road along the rift, are shown in Fig.1. Charges of up to 1,400 kg were used and placed on the lake beds at a depth of 25 m in Lake Rudolf and at the maximum available depth of 10 m in Lake Hannington. The two recording parties were equipped with identical systems consisting of eight 2 Hz seismometers laid out in a linear north-south array, and including a three component set near the centre point. Little improvement in signal/noise ratio was obtained by array processing methods but the first onset of shear waves was clearly visible on a product trace of the vertical and north (nearly radial) components of the three component set. The interpretation was, therefore, carried out using only the first arrivals of the compressional (P) and shear (S) waves. Straight lines were fitted to both the "first P" and "first S" arrivals from both shot points R_1 and R_2 at Lake Rudolf and from the shot point at Lake Hannington. Differences in the (extrapolated) overall travel times for the compressional wave data suggest that the arrivals from Lake Hannington and Lake Rudolf are from different refractors. Any attempt to construct delay time profiles assuming that the arrivals are from the same layer leads to improbable shot point delays. A simple interpretation in terms of horizontal layering was therefore made and is shown in Fig.2, an upper layer velocity for the surface lavas being assumed on the basis of short range data obtained in the vicinity of Lake Baringo. Almost all the arrivals from Lake Hannington appear to be refractions through rocks at no great depth beneath the lavas, those from Lake Rudolf refractions from a much deeper higher velocity layer. If an assumption is made about shear wave velocity in the top layer the interpretations based on P and S waves are consistent. The large standard error found for the velocity of the deeper layer is of interest and receives comment below. Also of importance, though no satisfactory explanation can be given, is the poor recording of the high velocity refractor from the shots in Lake Hannington.

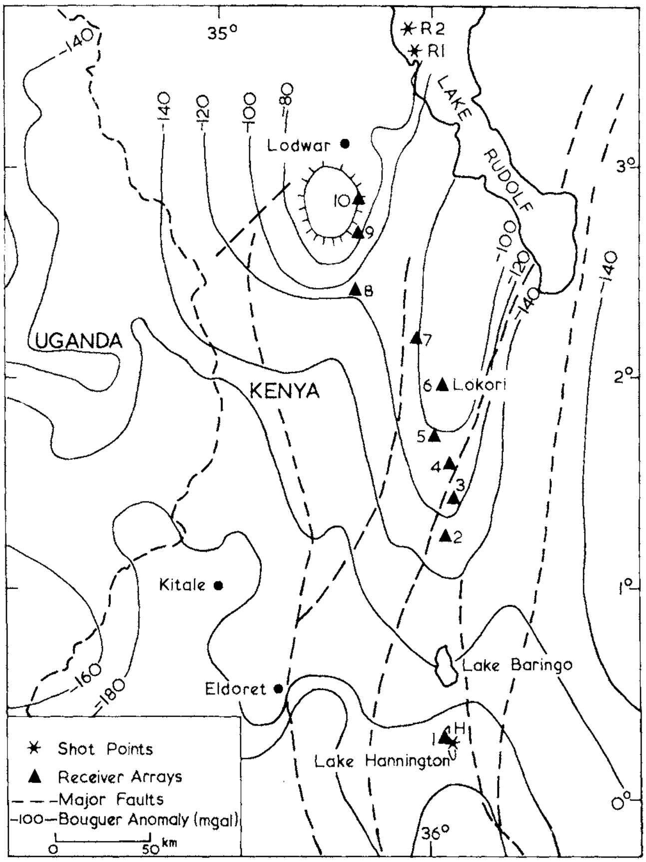

Fig.1. The northern part of the Kenya Rift showing shot points and seismic recording stations, and the Bouguer anomaly (after Khan and Mansfield, 1971).

INTERPRETATION

If for a moment this plane horizontal layer interpretation is accepted as a crude approximation to the structure and, as already mentioned, it does not seem possible to produce any other simple explanation of the data, then it appears that under the rift in this region there is, at a depth of about 20 km, a medium of high compressional and shear wave velocities. Teleseismic delays at Addis Ababa and Nairobi (Lilwall and Douglas, 1970) of the order of 2 sec relative to those at stations on shield areas suggest a minimum thickness of 100 km for this layer on the assumption that its contrast in velocity with normal mantle does not decrease with depth. One interpretation of the data, therefore, is that the asthenosphere is here cutting the lithosphere and penetrating the lower crust to within 20 km of the surface. This is consistent with the broad regional gravity low along the length of the rift zone. It is interesting to consider whether the refraction data shows

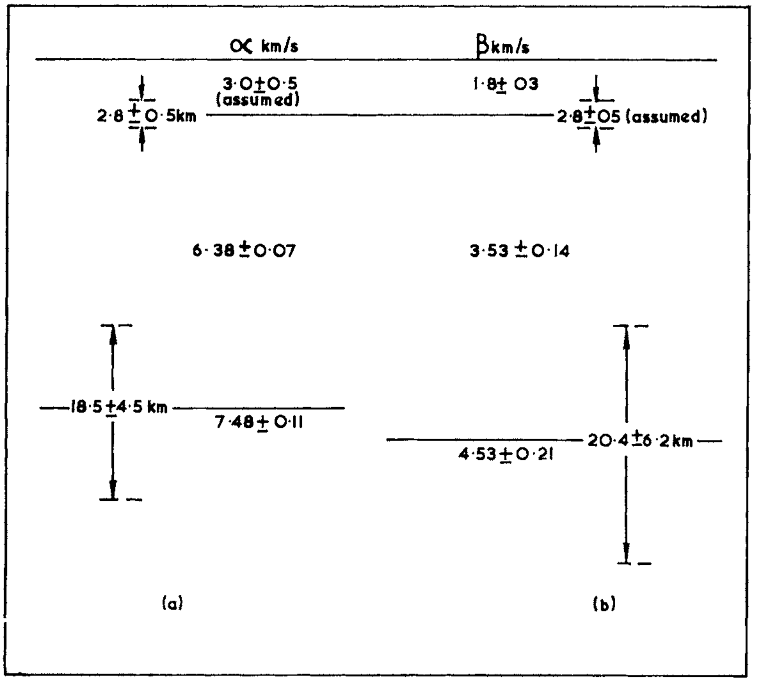

Fig.2. Interpretation of seismic refraction data in terms of plane parallel layering for (*a*) P waves, and (*b*) S waves.

any indication of the presence of a major intrusion at quite shallow depth which would similarly account for the axial gravity high. Taking into account the fact that a 6.3 km/sec velocity was obtained for arrivals from shots fired in Lake Baringo and recorded from short range out to 88 km it appears that rocks with about this or a slightly higher velocity occur directly beneath the volcanics, at a depth of some 3 km. Velocities of 6.3–6.4 km/sec are high for a normal continental area and, given the geological situation, may be explained as a result of crustal intrusion, perhaps in the form of dykes. It is interesting to note that first arrivals from Lake Hannington at stations 4, 5, 6, 7 are early for crustal arrivals, though only the arrival at station 5 seems early enough to be a refraction from the high velocity layer. Reduced travel times for the previously calculated crustal velocity of 6.38 km/sec are shown in Fig.3a. The fact that these early arrivals are observed at stations on the axial gravity high leads one to ask whether an axial basic intrusion might not be acting as a refractor of velocity intermediate between those of the 6.4 and 7.5 km/sec layers already postulated. The seismic data above are clearly inadequate to establish the presence of such a layer, but the plot of delay times shows that rocks with a 6.7 km/sec velocity could be present at a depth of about 10 km giving travel times consistent with the data. The velocity of 6.4 km/sec obtained for shots from Lake Hannington out to stations 8, 9 and 10, for which the travel path is off the axial gravity anomaly, is perhaps an indication of the laterally limited extent of the intrusion, though the velocity is still high for normal crust.

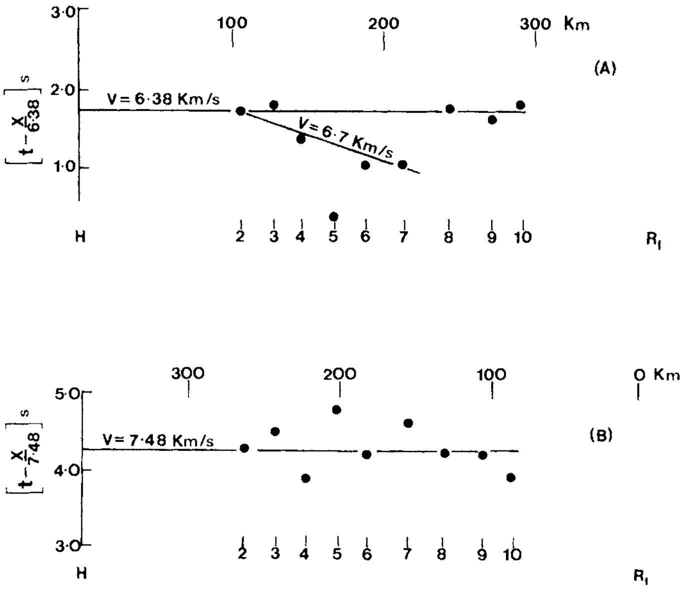

Fig.3. Reduced travel times for arrivals from (A) Lake Hannington, and (B) Lake Rudolf (shot point 1).

If we can think of crustal density in general decreasing radially outward and upwards from a massive basic intrusion at depth beneath the axial gravity anomaly and reaching the normal value for continental crust towards the margins of the rift where the degree of intrusion is supposed minimal, then the available density contrast is sufficient to at least contribute substantially to the anomaly.

The contrast is likely to be 0.1 g/cm^3 or greater, the range of gravity is around 50 mgal and the depth to the high velocity layer is 20 km. However, lateral variations in density in the crust of this magnitude and on this scale could not give rise to delay time variations from the 7.5 km/sec layer of more than 0.15 sec and so cannot account for the observed variations which are up to ± 0.5 sec. Also reasonable velocity variations in the high velocity layer, e.g., ± 0.2 km/sec over a distance of 30 km, can only give rise to changes in delay time of ± 0.1 sec. The observed delay times, as can be seen from the reduced travel times plotted in Fig.3b, change in a more or less random way along the seismic line and show no relationship to the changes in Bouguer anomaly.

CONCLUSION

Taken together the gravity and seismic evidence, therefore, appears to be consistent with the presence of an axial intrusion of basic magma rising to perhaps something less than 10 km below the surface. Above this the crust is partially intruded, perhaps passing laterally, away from the rift, into normal continental crust. At a depth in the region of 20 km there is a highly irregular, possibly in part gradational contact with the 7.5 km/sec layer which both teleseismic and gravity data suggest to be of the thickness of the normal

lithospheric plate. The structure proposed combines features of the crustal models presented by Khan and Mansfield (1971) and Searle (1970) and is in fact very similar to that given by Baker and Wohlenberg (1971).

When more gravity measurements have been made and the axial positive anomaly better defined it would be valuable to obtain further information on crustal velocities. This could best be done by shooting a number of seismic lines of moderate length both on and off the anomaly, preferably parallel to the rift axis. As a result of such measurements it should be possible to confirm or disprove the presence of an intrusion and obtain some information about its nature and geometry.

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RAYLEIGH WAVE PHASE VELOCITIES FOR THE PATH ADDIS ABABA–NAIROBI

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ABSTRACT

Knopoff, L. and Schlue, J.W., 1972. Rayleigh wave phase velocities for the path Addis Ababa–Nairobi. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 157–163.

Rayleigh wave phase velocities have been obtained for the path Addis Ababa–Nairobi in the period range 20–125 sec. The phase velocities are low over the entire range when compared with most other parts of the world. The phase velocities are the same as for the Great Basin of the western United States. Inversion has been made by the "hedgehog" method. Probably the mantle has an S_n velocity of between 4.25 to 4.45 km/sec extending to depths of perhaps 120–200 km below the crust. A thin high velocity veneer at the top of the mantle is not excluded by our data.

INTRODUCTION

Surface wave dispersion measurements taken to periods of the order of 100 sec or more have proved to be diagnostic of the properties of the upper mantle to depths of the order of $0.4\lambda_{mR}$ for Rayleigh waves and $0.25\lambda_{mL}$ for Love waves, where λ_m is the longest wavelength. The technique in common use today is the "two station" method, in which distant earthquake sources are used that lie close to the great circle extension of the path between two seismic stations. The seismic stations are most appropriately similarly instrumented with long period seismometers. The records should be on scale at both stations. That the two stations should span a relatively homogeneous geophysical structure is important only for the purposes of interpretation of the dispersion curves. The method of interpretation will be discussed below.

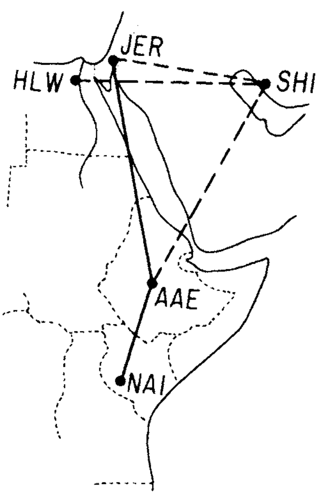
As applied in this laboratory, the data processing proceeds in several modular stages. The records are digitized at an interval sufficiently short to minimize aliasing; that is, the digitization interval is short compared with the period of significant power in the signal spectrum on the particular recording. For typical WWSSN long period records, this interval is usually 1 or 2 sec. The records are then filtered with a band pass filter whose center time is close to the group arrival time of the period in question. The filtered record is windowed to reject echoes and the resultant record is Fourier analyzed. The phases from such overlapping period bands are compared for the two stations and the phase

velocities are computed. The arbitrary integer in the phase velocity calculation is determined by making the phase velocity "reasonable" at long periods and extending the phase continuously into the short period regime. Rayleigh wave phase velocities obtained by this method, and taken to long periods, are catalogued for a variety of geographic regions by Knopoff (1972).

Earlier efforts at interpreting the structure of a given region using the dispersion of surface waves have attempted to make use of the simultaneous observations of both Love and Rayleigh waves. The possibilities of structural fit to the observations are reduced, if both are used in the interpretation, compared with the case of the interpretation of only one of these dispersion curves. Unfortunately, Thatcher and Brune (1969) and Boore (1969) showed that a systematic bias exists in the Love wave observations, which would obviously prejudice any attempt at interpretation; we have no way to get around this difficulty at this time. We report only the results of measurement and interpretation of fundamental mode Rayleigh wave phase velocities in this paper.

ANALYSIS AND INTERPRETATION

We have analyzed the seismic records at WWSSN stations Addis Ababa (AAE) and Nairobi (NAI) in an attempt to obtain useful dispersion data for the purpose of structural interpretation of the East African region (Fig.1). The records from four earthquakes were analyzed at these two stations by the techniques above. These events, sufficiently close to the great circle path between AAE–NAI (Fig.2) are given in Table I. The three Love wave phase velocity curves are not reported here for reasons given above. Of the three Rayleigh



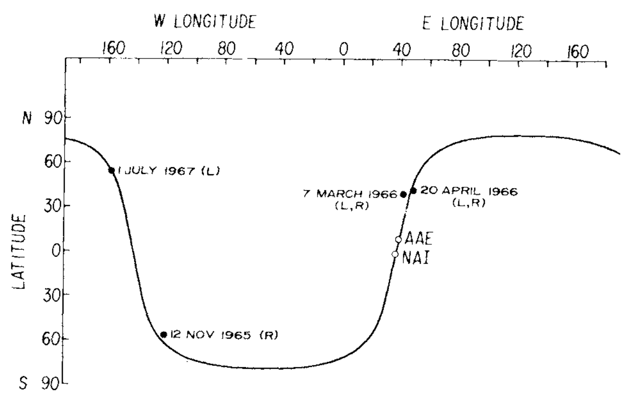


Fig.1. Regional sketch map showing Worldwide Standard Seismographic Network (WWSSN) stations and phase velocity profiles.

Fig.2. Great circle path through AAE–NAI and location of the four earthquakes whose records were analyzed.

TABLE I

Earthquakes recorded at Addis Ababa and Nairobi

Date	Location	Wave
12 November 1965	South Pacific	R
7 March 1966	Turkey	L, R
20 April 1966	Turkey	L, R
1 July 1967	Aleutian Islands	L

wave phase velocities, the South Pacific event gives a profile reversed from those of the two Turkish events. The three curves are in good agreement over their common range of periods. The average Rayleigh wave phase velocities thus obtained are shown as the solid circles in Fig.3, for the period range 20–125 sec.

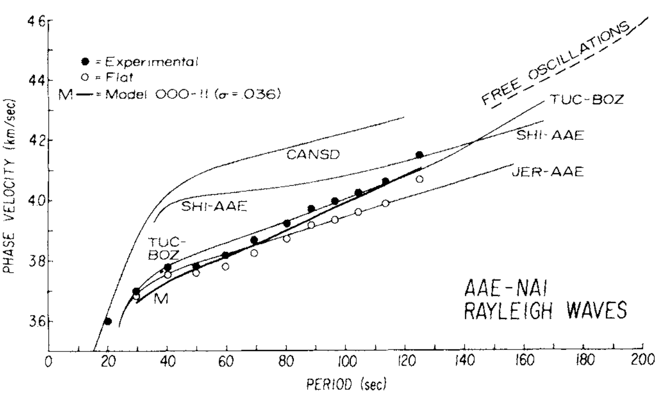


Fig.3. Rayleigh wave phase velocities observed for the path AAE–NAI (solid circles) and reduced to a flat earth (open circles). A theoretical flat-earth model with $\sigma = 0.036$ km/sec is shown (heavy line). Other Rayleigh-wave phase velocity profiles for comparison.

For purposes of comparison, the Rayleigh wave phase velocities for other selected station pairs are also given in Fig.3. These are summarized, along with others, by Knopoff (1971). The observed Rayleigh wave phase velocities for the path AAE–NAI are comparable to those obtained by Dr. N.N. Biswas (1971) of this laboratory★ for the WWSSN path in the Basin and Range province of the western United States, Tucson (TUC)–Bozeman (BOZ). Mr. A.A. Fouda, also of this laboratory, has obtained phase velocities for Rayleigh waves for WWSSN paths near and across Arabia, namely Shiraz (SHI)–Jerusalem (JER), SHI–Helwan (HLW), SHI–AAE and JER–AAE (Fig.1). The profile JER–AAE crosses the Red Sea and has phase velocities comparable to those for AAE–NAI, although they are somewhat lower at long periods. The three trans-Arabian profiles have similar phase velocities among themselves; only SHI–AAE is shown here.

★ Presently at University of Alaska

Significant differences are found in the intermediate period band (40–100 sec) when the profile SHI–AAE is compared with the profile AAE–NAI. The Canadian Shield profile (Brune and Dorman, 1963) gives extraordinarily high velocities compared with AAE–NAI. We have also indicated the phase velocities corresponding to high order spheroidal modes of free oscillation of the earth. These latter should represent global averages, in some sense, and not be representative of any particular region.

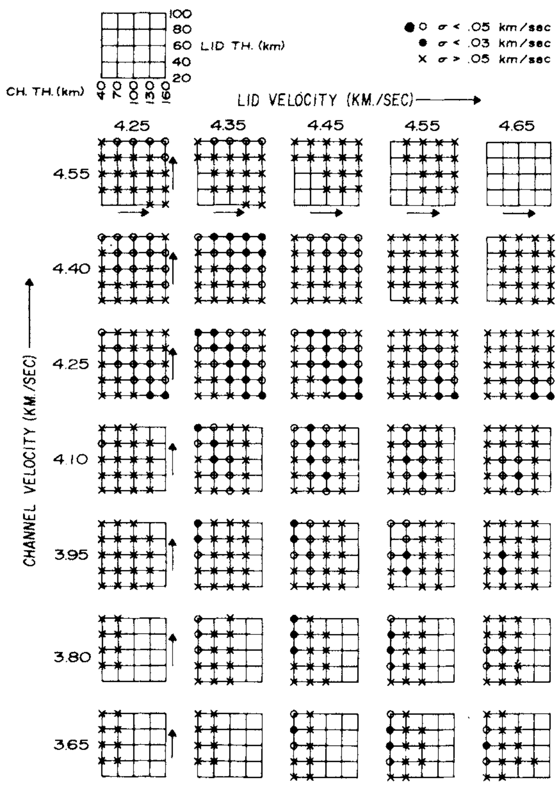
The interpretation of the phase velocity curve is accomplished in several steps. First, we reduce the values to take account of sphericity. We use the correction of Bolt and Dorman (1961). The flat-earth equivalent phase velocities are shown as open circles in Fig.3. These reduced values are interpreted by a computer program designed by Keilis-Borok and Knopoff called "hedgehog". Hedgehog is a search in a multi-dimensional parameter space for solutions to a given problem. We parameterize the earth structure: in the search described in Fig.4, only four degrees of freedom in the search were permitted, namely possible lid and channel thicknesses, and lid and channel S-wave velocities. The crust is assumed to have known properties, as well as the mantle below a possible low velocity channel; the P wave velocities and densities are assumed either to have known properties in the channel and lid, or to be correlated to the S wave properties in these regions. The program requires that a point in the parameter space be found which is "successful", i.e., it gives a phase velocity which fits the flat-equivalent observations to within certain accuracies. Usually, our accuracy criteria are that no single calculated phase velocity should differ from its corresponding flat-observed value by more than 0.1 km/sec and that the r.m.s. deviation of phase velocities for all points should not exceed a certain value σ. In the two experiments of Fig.4, σ was taken to be 0.05 km/sec and 0.03 km/sec.

From the starting point, the program searches nearest neighbors for additional successful points, until the entire singly connected space of successful points has been explored. The search is diverted into the space of successful points either because the search has taken the program into a region of unsuccessful solutions or because the bounds of the parameterization have been reached.

The crustal structure and value of P_n for the AAE–NAI inversion was taken from the refraction results of Griffiths et al. (1971). We noted the observations of Dopp (1964), but did not use them. The recent observations of Searle and Gouin (1971) were not available to us at the time of computation. The lower mantle was fixed to have step discontinuities at 400 km and 600 km.

The structures explored are shown in the 4-dimensional representation of Fig.4. The unsuccessful structures for $\sigma = 0.05$ km/sec are indicated by crosses. The successful structures at this level are the circles. The successful structures at the level $\sigma = 0.03$ km/sec are indicated by solid circles and are, of course, a subset of the $\sigma = 0.05$ km/sec collection. Several points were not explored due to termination of the computer run; of these, most would appear to be predictably structures which would have been rejected, except for the following:

(80, 40, 4.55, 3.95)
(100, 70, 4.55, 3.80)
(20, 70, 4.55, 3.95)



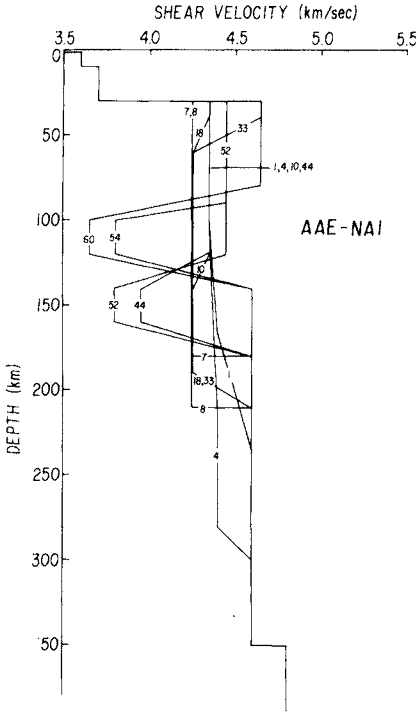


Fig.4. Four-dimensional space of solutions explored by "hedgehog". Successful solutions are indicated by circles. Each small grid has, as coordinates, channel thickness and lid thickness (see inset above). Each column of grids corresponds to a different lid velocity while each row of grids corresponds to a different channel velocity.

Fig.5. Representative successful structures from the parameterization with criterion $\sigma < 0.03$ km/sec.

in the order of variable (lid thickness, channel thickness, lid velocity, channel velocity). We have no information about these three structures. Several representative members of the set of successful cross sections are plotted in Fig.5.

The nomenclature of "lid" and "channel" is for identification only. These names identify the topmost two layers of our parameterization of the mantle. In some cases, especially in the upper left hand corner of Fig.4, the S wave velocity in the lid can be less than the channel velocity, thereby implying a positive velocity gradient and the absence of a literal channel.

Since the lid is taken to be homogeneous in these models, we identify the velocity in this region with S_n for the purposes of the subsequent discussion.

As we believe our experimental data to be more precise than $\sigma = 0.03$ km/sec, we focus our attention on the solid circles of Fig.4. First, we consider the results:

(20, 130, 4.25, 4.25)
(20, 160, 4.25, 4.25)

These cases indicate that a single upper mantle layer 150 to 180 km thick with a uniform velocity of 4.25 km/sec will fit the observations. Cases such as (70, 100, 4.25, 4.25) which would seem to fit between the above two successes were in fact rejected because of some differences among the P wave velocity and density structures. We do not believe that this can remove the possibility that a thick upper mantle layer with velocity around 4.25 km/sec and without a significant velocity gradient can fit the observations.

The solutions:

(100, 70, 4.35, 4.40)
(100, 100, 4.35, 4.40)
(100, 130, 4.35, 4.40)
(100, 160, 4.35, 4.40)

indicate that we are unable to provide resolution for the S-wave velocity at depths greater than 200 km below the surface with our data.

What these solutions do show is that the upper mantle has an extensive region of material with unusually low S-wave velocity. Many of the solutions seem to show that a region of from perhaps 120 km to 200 km in thickness has a velocity of S_n of between 4.25 to 4.45 km/sec with little or no velocity gradient in it. In this group of solutions, zero velocity gradients are obtained for S_n velocities 4.25 and 4.35 km/sec. If S_n is as high as 4.45 km/sec, a low contrast low velocity channel with velocity 4.25 km/sec is the channel with least velocity contrast to lid.

On the basis of our data alone, we cannot reject the possibility that material with even lower S wave velocity can exist at depth. However, in each of these cases, this anomalous low velocity is found in a channel. Some narrow channels with channel velocity 4.10 km/sec are found even for lids 60 to 100 km thick with S_n velocity 4.35 km/sec.

If the S_n velocity is high, e.g., 4.55 or 4.65 km/sec, the lids are relatively thin and the channel velocities are extremely low. For example, with a channel velocity of 4.25 km/sec, we could put a veneer of 20 km of 4.65 or 4.55 km/sec lid material above it; we would be unable, with our data, to resolve this model from one without the veneer. If we lower the channel velocity, we can increase the amount of high velocity material above it. Evidently, our results require a low average S wave velocity in the upper mantle, without serious regard to the details of the averaging.

To return to Fig.3, we can conclude that the Basin and Range province of the western United States similarly has low velocity material below the crust. It is not inconceivable that the western United States and the East African Rift have similar gross mantle structure. The Arabian peninsula to the north of AAE–NAI, on the other hand, has a thick zone of high velocity material beneath the crust and overlying the low velocity channel. The generally high phase velocities for the Canadian Shield indicate that the crust is underlain by high velocity material to great depth, with the channel possibly absent completely. The profile AAE–NAI gives some structures that make it look as if the lid to a high contrast channel has been thinned to, or almost to, vanishing.

Since the profile JER–AAE traverses only in part a patently anomalous zone, the

inverted structure is clearly mixed. If the continental parts are much like those under the Arabian profile SHI–AAE, then the structure under the Red Sea must be even more anomalous than that which we have described for AAE–NAI. But the discussion of the profile JER–AAE falls outside the scope of this paper; that will be made by Mr. Fouda elsewhere.

Finally, we must comment about lateral inhomogeneities. We have interpreted the phase velocity curves using theoretical structures for a horizontally layered half-space. Certainly, at some distance to either side, high velocity material must adjoin the anomalous East African rift zone. This would give us the configuration of a three dimensional wave guide in the form of a horizontal, low velocity conduit surrounded by higher velocity material on the sides as well as below; the higher velocity material on the flanks is also vertically inhomogeneous. The effect of the higher velocity material to either side must be to raise the observed phase velocities from those for a half-space with rift stratification. The corrections are probably smaller at the short wavelength end of the spectrum than at the long. This can only mean that our estimates of structure in Fig.4 and 5 have average S-wave velocities that are too high. At this time, we give no estimate of this error. Hence, our failure to take into account the finite lateral extent of the rift zone does not weaken our conclusion that the S-wave velocities in this region are anomalously low.

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THE STRUCTURE OF EAST AFRICA USING SURFACE WAVE DISPERSION AND DURHAM SEISMIC ARRAY DATA

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ABSTRACT

Long, R.E., Backhouse, R.W., Maguire, P.K.H. and Sundarlingham, K., 1972. The structure of East Africa using surface wave dispersion and Durham seismic array data. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 165–178.

As a background to the discussion of the array data some results of studies on teleseismic P-wave delays and surface wave dispersion within the rift zone are presented. This work uses data from permanent stations. The deep structure of the Gregory Rift is subseqently discussed using data from the Kaptagat array station installed by the University of Durham in Kenya in 1968. A compressional velocity model for the Gregory Rift is presented.

INTRODUCTION

Away from the rift zone, Africa has a structure typical of stable shield areas. Gumper and Pomeroy (1970) have derived a mean structural model for Africa from the dispersion of Rayleigh waves along paths outside the rift zone. This model, which will be referred to as the AFRIC model is a minor modification of the CANSD model for the Canadian shield (Brune and Dorman, 1963). Similar models have been derived from surface wave dispersion studies in southern Africa by Bloch et al. (1969) (Fig. 1).

In contrast, a refraction line in the Gregory Rift reported by Griffiths et al. (1971), shows a 20-km thick layer of 6.4 km/sec overlying a 7.5 km/sec layer presumed to be anomalous mantle material. Such an anomaly is to be expected from the failure of S_n to propagate across the rift zone north of the equator, which has been taken by Gumper and Pomeroy (1970) to indicate "a gap in the lithosphere".

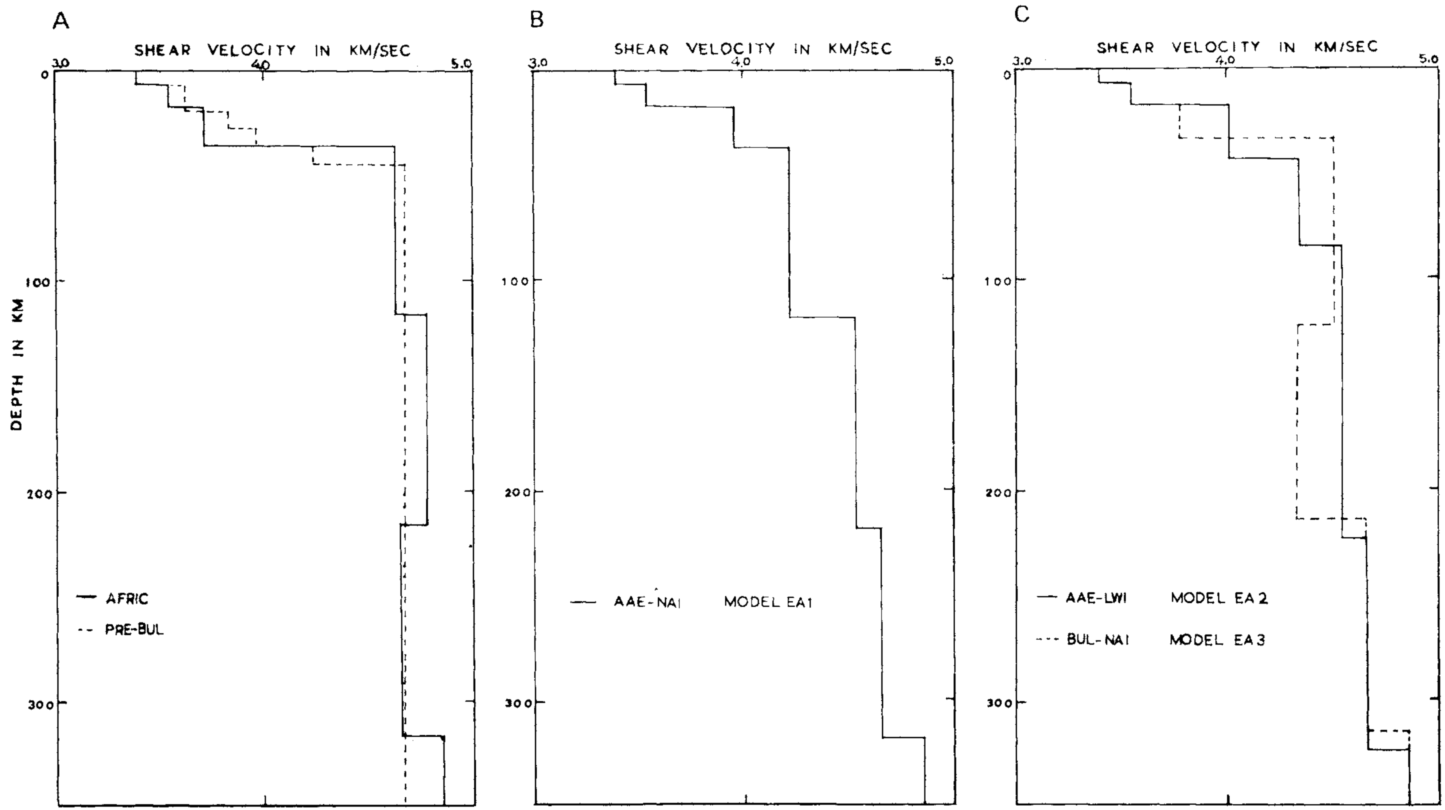

Fig. 1. Optimum models (B, C) for the various inter-station paths compared with (A) the AFRIC model of Gumper and Pomeroy (1970) and a model for the Pretoria-Bulawayo path from Bloch et al. (1969).

THE GENERAL STRUCTURE OF EAST AFRICA

Seismic data from permanent stations have been used to give a regional structure for East Africa and to locate regions of mantle anomalies.

Teleseismic P wave delay

Station residuals show the WWSS stations at Addis Ababa and Nairobi to have large positive delays (Herrin and Taggart, 1968; Lilwall and Douglas, 1970), whilst Bulawayo has a small negative delay (typical of shield regions).

Such station residuals may contain source and non-typical path delays leading to uncertainties in their interpretation. The direct measurement of relative delay between a pair of stations largely (though not completely) removes such effects, so that relative delay represents a true measure of the crust and upper mantle differences between stations (Long and Mitchell, 1970).

Sundaralingam (1971) has measured delays at Addis Ababa, Nairobi and Lwiro relative to Bulawayo using events in the distance range 25°–90°. Corrections are applied for altitude, differences in epicentral distance and angle of emergence using Herrin's tables, (Herrin et al., 1968). The results are shown in Table I. No significant variations with azimuth or epicentral distance have been found.

It is evident from surface wave dispersion studies (e.g., the PRE–BUL model of Fig. 1) that Bulawayo lies on a shield crust and mantle typical of Africa as a whole. These relative delays are therefore a direct measure of the differences between the upper mantle beneath the rift and shield. The delays indicate the existence of a substantial low velocity zone in the upper mantle beneath the Eastern Rift (as sampled by NAI), while some anomalous material appears to be present but less extensive beneath the Western Rift (LWI)

It is noteworthy that the delays associated with Addis Ababa and Nairobi are similar, and are close to the large relative upper mantle delay of 2.5 ± 0.4 sec calculated by similar methods for Iceland relative to the shield station of Kiruna in Sweden (Long and Mitchell, 1970). This implies that the mean compressional velocity of the upper mantle beneath the Gregory Rift is similar to that beneath Iceland and therefore presumably to that beneath the Mid-Atlantic Ridge.

TABLE I

P wave teleseismic delay times relative to Bulawayo with 95% confidence limits

Locality	Delay
Addis Ababa	2.7 ± 0.3 sec
Nairobi	2.3 ± 0.3 sec
Lwiro	1.1 ± 0.3 sec
Eastern Rift station mean	2.5 ± 0.3 sec
Mantle delay for Iceland relative to Kiruna	2.5 ± 0.4 sec

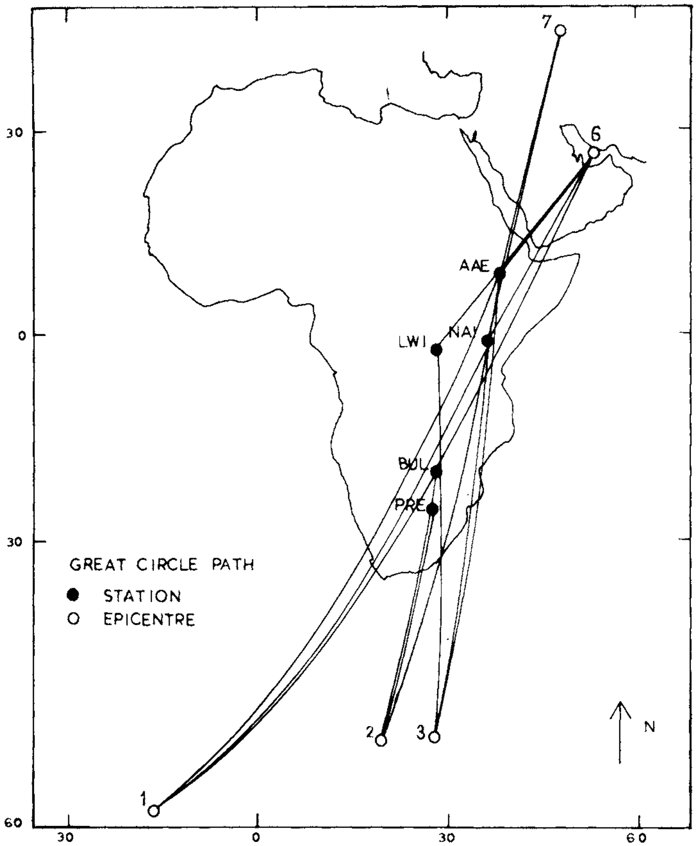

Fig. 2. Events and stations used in the dispersion measurements for Fig. 3. Lines are the great circle paths from the events to the stations.

Surface wave dispersion along the rift zone

Sundaralingam (1971) has studied the dispersion of Rayleigh waves travelling between the permanent stations of Addis Ababa (AAE), Nairobi (NAI), Lwiro (LWI) and Bulawayo (BUL). Inter-station phase velocities were measured for events close to the great circle paths through these stations (Fig. 2) using Sato's (1955) fourier transform method.

With the exception of the AAE–LWI path which was only sampled from the north, the paths were reversed giving identical dispersion. It is therefore unlikely that the curves are strongly affected by refraction or extra-inter-station path effects.

Fig. 3 compares the principle data from this work with dispersion for the AFRIC model. Two immediate observations may be made. First, the curves tend to merge at shorter periods indicating some uniformity of the crust over Africa as a whole. Second, compared with the AFRIC model there is a significant reduction in phase velocity at higher periods, indicating anomalously low upper mantle velocities under the rift zone.

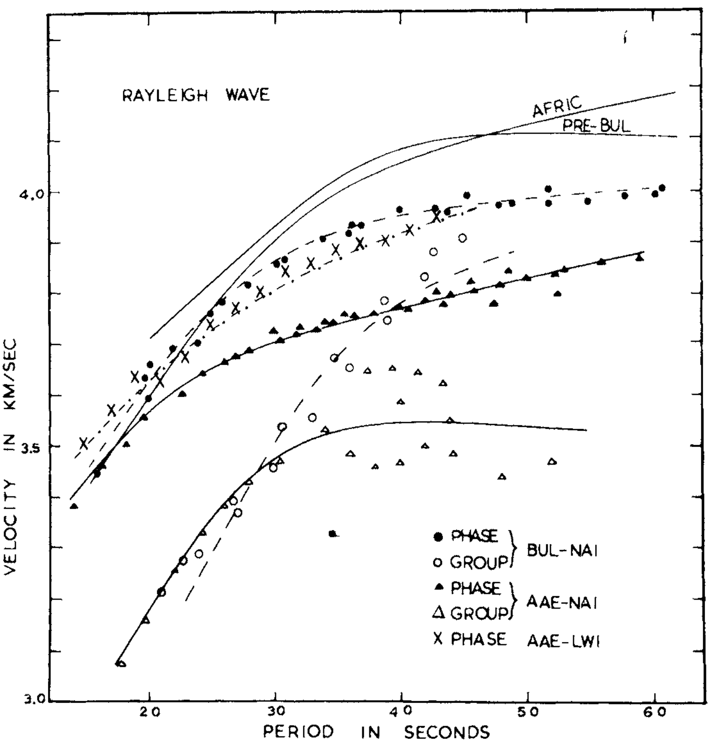

Fig. 3. The principal Rayleigh wave dispersion data. The lines through the points refer to the corresponding models of Fig. 1.

Sundaralingam (1971) has derived models from these data within the following context.

(1) The period range from 15 to 60 sec provides information on structure between 17 and 120-km depth. Parameters for layers outside this range have been assumed at the AFRIC values.

(2) Partial derivatives show that compressional velocity and density of the various layers do not significantly control the dispersion. Thus compressional velocity is calculated as 1.74 times the shear wave velocity, and densities are generally assumed at AFRIC values but constrained to be consistent with the regional gravity data.

(3) Backus and Gilbert (1968) have shown that there is a fundamental limit to the detail of earth structure which may be derived from the gross earth data. Der et al. (1970) have applied a similar argument to the surface wave inversion problem. They have shown that the inherent resolution of the data can be expressed as a minimum thickness of a zone over which velocity may be determined to a given accuracy. Thus in the present study, thicknesses have been chosen so that shear velocities may be determined to an accuracy of about 0.1 km/sec. As a consequence, the layer boundaries are not structural boundaries and the shear velocities are to be regarded as firmly determined means over the layer thick-

nesses. Structure within a layer is not ruled out, but a firm determination of such structure is beyond the resolution of the data.

(4) The one exception to (3) was that the lower "crustal" layer thickness was allowed to vary to obtain some indication of the thickness of the crust. Thicknesses are not, however, well determined by this data.

Within this context statistically best fitting shear velocity models for the various paths have been determined by a process of optimisation, confirmed by exhaustive mapping. The principal models are shown in Fig. 1. In these models, only the velocities for the lower two crustal layers and the uppermost mantle layer were determinate. The range of uncertainty for the velocities of these layers is ± 0.15, ± 0.10 and ± 0.10 km/sec respectively in each model.

There is a firm indication of anomalously low velocity material in the topmost mantle. This zone is well developed in model EA1 for the AAE–NAI path, but is less extensive along the other paths.

The AAE–LWI path is probably a mixed path which can be split into three sections of approximately equal length: the eastern flanks of the Western Rift, an apparently undisturbed region between the eastern and western rifts and the western flank of the Ethiopian Rift. If the Ethiopian section is assumed to have AAE–NAI dispersion and the central section AFRIC type, then the dispersion of the Western Rift section is similar to that of the path as a whole. Model EA2 is thus a useful indication of the structure of the Western Rift.

Again, the NAI–BUL dispersion may be considered as a mixture of the dispersions for the AAE–NAI and for the PRE–BUL paths (Fig. 1). Thus it may be considered to contain a transition from the structure of the Eastern Rift to that of Africa as a whole. This dispersion is consistent with EA1 structure extending from Nairobi south-southwest across Tanzania to the Western Rift with normal structure for the remainder of the path.

We conclude that the main mantle anomaly extends along the Eastern branch of the rift system and that while some mantle anomaly probably exists beneath the Western Rift it is likely to be much less extensive. The precise structure of the latter is not resolved by the dispersion data.

Similarly, since the layer boundaries are not structural boundaries, the position of the Moho cannot be determined by these data alone. It is noteworthy that the velocities of the upper three layers in the various models are close to the AFRIC values. Whilst parameters for the topmost layer are assumed, mean velocities for the two lower "crustal" layers are determined to ± 0.15 and ± 0.1 km/sec respectively. The velocity for the second layer is remarkably similar to that of the AFRIC model in all cases. For the third layer, the determined mean velocity for the NAI–BUL path is not significantly different from the AFRIC values but there is a significant difference from AFRIC along the AAE–NAI and AAE–LWI paths.

Although this could indicate some modification of the lower crust, such a high velocity third layer could be generated if normal AFRIC crust was underlain by a thin zone of material with near-normal sub-Moho velocity forming a lid to an ultra-low-velocity zone in the topmost mantle. The lid cannot be resolved and consequently the mean velocities of

the lower crustal layer and the topmost mantle layer, together with the thickness of the lower crustal layer, would be increased. These surface wave data therefore do not require a crust of the type suggested by Griffiths et al. (1971) for the axis of the Gregory Rift, but are more consistent with normal AFRIC crust probably with near-normal sub-Moho velocities extending over the rift zone.

This is not in conflict with the refraction result indicating crustal modification along the Gregory Rift (Griffiths et al., 1971) since the AAE–NAI path lies generally east of the Gregory Rift and consequently is unlikely to sample the anomalous structure which gravity data suggests is restricted to the Rift axis (Khan and Mansfield, 1971). Surface wave data allows the possibility of near-normal crust and topmost mantle within some 50 km of the Rift axis.

SEISMIC ARRAY DATA

Three temporary seismic array stations have been set up in East Africa. Two L-shaped arrays of ten short-period instruments at 1 km spacing were installed at Kaptagat in Kenya and at Kakumiro in Uganda. The third array of three radio-linked, short-period instruments was sited at Murchison Falls.

The purpose of the network is to investigate the seismicity and the crust and upper mantle structure using a wide range of array techniques. Some preliminary results on the crust and upper mantle structure associated with the Gregory Rift are presented using data from the Kaptagat array, which is situated some 10 km west of the Elgeyo escarpment on the western flank of the Gregory Rift.

Apparent velocities of regional earthquake arrivals

Regional events have been located using azimuths measured by the array and distances determined mainly from P to S times. Of these events a group with well defined onsets has been selected (Fig. 4) and their apparent velocities measured by least-squares fitting of a straight line to the first arrivals. Epicentres are shown in Fig. 4 and their apparent velocities plotted as a function of distance in Fig. 5. Focal depths have not been taken into account so Fig. 5 cannot be immediately interpreted as a velocity–depth model.

The events fall into two groups:

(1) those events from the western sector (solid circles of Fig. 4 and 5) whose first arrivals show velocities of between 5.7 and 6.4 km/sec for epicentral distances less than about 200 km and 7.9 ± 0.3 km/sec for greater distances.

(2) those events whose paths pass along or across the Gregory Rift (triangles in Fig. 4 and 5) which show velocities of 7.1 ± 0.3 km/sec.

The variation in velocity for the near events of group (1) may be due to an increase of velocity with depth within the crust revealed by varying focal depth. These are crustal events with a mean focal depth of about 15 km. For greater distances the apparent velocity of 7.9 km/sec is almost certainly to be associated with head wave arrivals from the

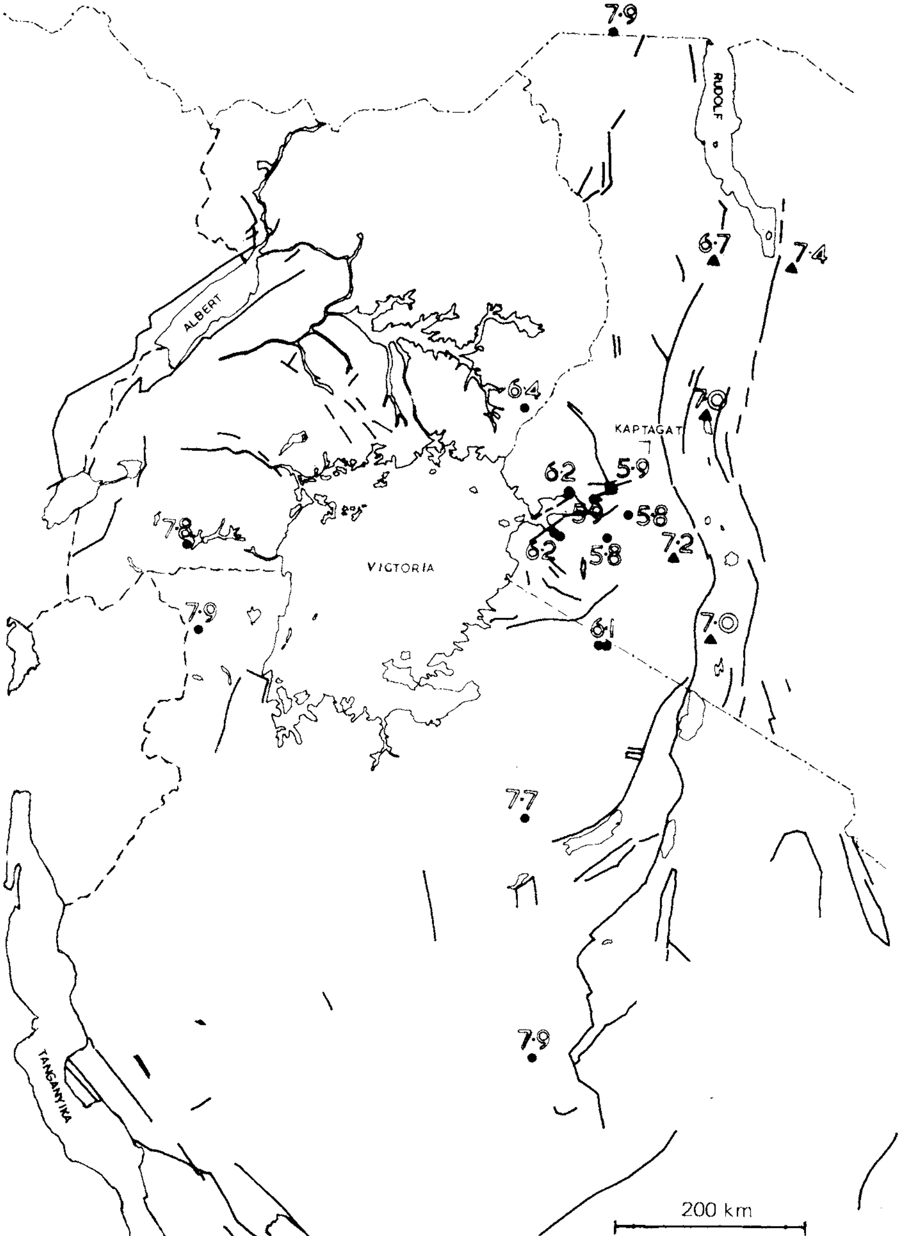

Fig. 4. Epicentres used in the crustal velocity measurements. The figures are apparent velocities in km/sec of first arrivals measured at Kaptagat.

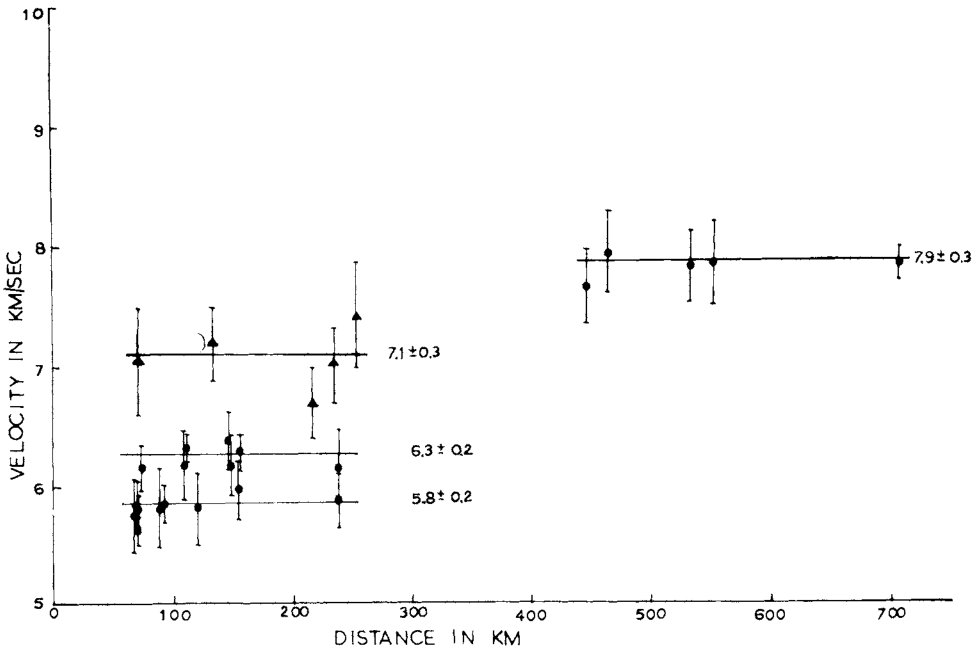

Fig. 5. Preliminary apparent velocity versus epicentral distance plot for the data of Fig. 4. Error bars show 95% confidence limits.

Moho. The absence of a continuous azimuthal variation suggests a horizontally uniform structure for the crust immediately to the west of the Gregory Rift. The apparent velocities are therefore probably realistic measures of actual crustal and sub-Moho velocities.

These velocities are typical for normal shield crust and mantle. Their values are similar to velocities determined by refraction using earthquake sources in the Transvaal (Wilmore et al., 1956; Gane et al., 1956; Hales and Sachs, 1959) which were between 6.0 and 6.3 km/sec for P_g and between 7.96 and 8.2 km/sec for P_n. These data therefore provide evidence for normal crust and topmost mantle between the two branched of the rift system, in agreement with the satisfactory propagation of S_n between the Gregory and Western Rifts (Gumper and Pomeroy, 1970).

In contrast, events of group (2) have sampled the anomalous structure along the axis of the Gregory Rift. Presumably the focal depths of all these events have been such as to give paths through the anomalous upper mantle which from seismic refraction work is detected at a depth of some 20 km. The discrepancy between the apparent velocity of 7.1 km/sec and the 7.5 km/sec determined from the refraction data (Griffith et al., 1971) may result from structural complexity. The array data confirms crustal modification along the Gregory Rift.

Fig. 6 shows a compressional velocity model for the crust in the vicinity of the Gregory Rift. The observation of 7.9 km/sec on the flanks to the west suggests the possibility of a

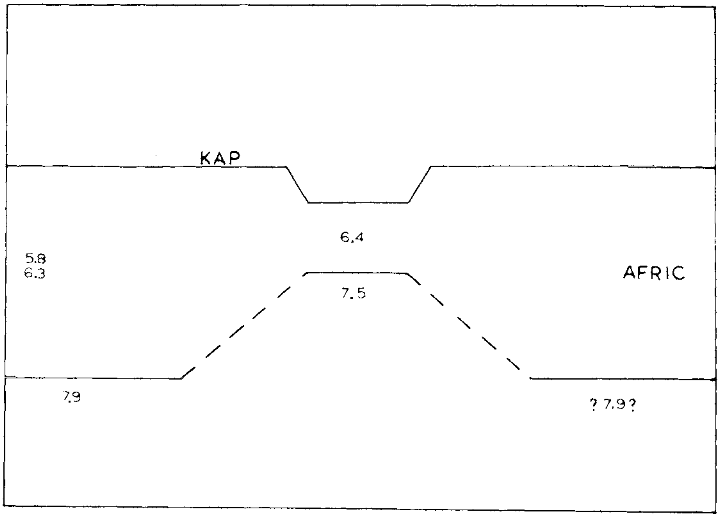

Fig. 6. Compressional velocity model of the crustal structure across the Gregory Rift.

similar sub-Moho velocity along the eastern flank. This would be consistent with the normal velocity lid above the low velocity zone which has been suggested earlier to account for the apparently high velocity of the third layer in model EA1 for the AAE–NAI path.

Apparent velocities of teleseismic arrivals

Unlike regional events, the apparent velocities of first arrivals from teleseismic events show a marked continuous variation with azimuth. These data are summarized in Fig. 7. The lack of continuous azimuthal variation in the apparent velocities of regional events and the general consistency of their velocities, make it highly unlikely that the azimuthal variation shown in Fig. 7 results from local perturbation in subsurface geology or from errors in array geometry. There is therefore a clear indication of some structural complexity beneath Kaptagat which is apparently not encountered by the crustal and sub-Moho arrivals from regional events.

The azimuthal dependence shown in Fig. 7 might be expected for some sloping boundary as has been shown to exist beneath other arrays (Niazi, 1966, Greenfield and Sheppard, 1969; Corbishley, 1970) but such a large azimuthal variation is unusual. Rather than discuss the individual boundaries it is more instructive to consider the wedges or prisms formed by such boundaries. These may be collectively considered as a single resultant prism. The many properties of such a prism are well known from geometrical optics (e.g., Heath, 1887). In all cases a low velocity prism in a higher velocity medium will deviate rays away from the refracting edge, thereby increasing the apparent velocities of arrivals from the thicker side and decreasing those from the thinner side (Fig. 8). Detailed analysis suggests

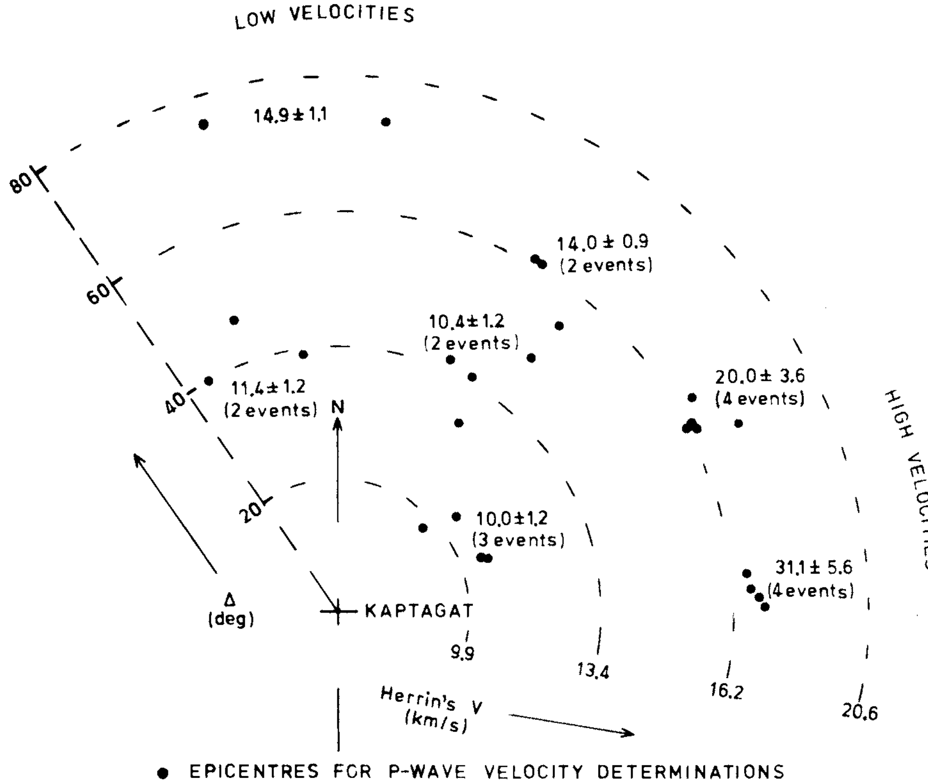

Fig. 7. Summary of apparent velocities in km/sec of teleseismic P waves measured at Kaptagat. Velocities are means for small groups of events, with 95% confidence limits. Expected velocities, from Herrin (1968), are given at 20° distance intervals.

that the observed apparent velocity pattern can be generated by a low velocity prism thinning westward.

The existence of normal crust in proximity to the thinned crust of the rift axis could indicate a sloping Moho, with possibly a gradation in velocity from 7.9 km/sec on the flanks to 7.5 km/sec along the axis. Again, the refraction work of Griffiths et al. (1971) yielding a 6.4 km/sec crust along the axis of the rift could suggest some modification of the crust itself. Both would yield a high velocity wedge thinning westward which could not explain the observed teleseismic result. This is summarized by the axial positive gravity anomaly (Khan and Mansfield, 1971) which suggests that the net effect of the various crustal modifications is a wedge of high density and therefore presumably high velocity, thinning towards the flanks. We may thus conclude that the principal cause of the teleseismic anomaly does not lie within the crust nor is it a result of Moho compexity.

The anomaly may be explained if the low velocity zone in the upper mantle thinned to the west. As shown in Fig. 6 the low velocity zone which presumably cuts into the crust along the axis of the Gregory Rift must deepen westward to give way to normal sub-Moho

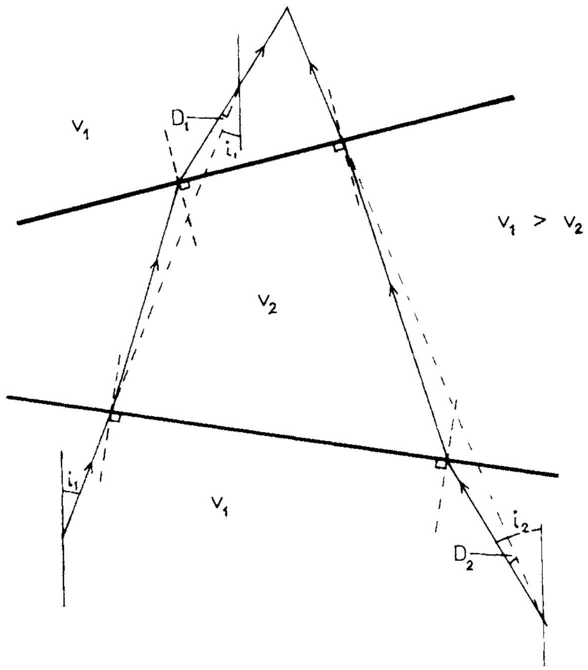

Fig. 8. The effect on apparent velocity of a low velocity prism set in a high velocity medium. The apparent velocity, $v_1/\sin i$, which is a constant for a ray in a spherically uniform earth is modified by the deviation D, to $v_1/\sin (i \pm D)$. The sign for D depends on the direction of approach of the incident ray to the vertical. The deviation is in the same sense irrespective of the angle of incidence of the arrival.

material. It is likely, therefore, that the teleseismic anomaly is a direct measure of the attenuation of this anomalous zone away from the Rift axis.

An indication of velocity within the anomalous material arises from a simple consideration of the zone as a prism with sharp velocity contrasts at its boundaries. The absence at Kaptagat of a significant variation of delay time with azimuth or distance, limits the wedge angle according to the velocity. Wedges fitting the $dT/d\Delta$ variation are found to require angles above this limit unless the velocity of the material is less than about 7.0 km/sec.

A similar anomaly in apparent velocity would be observed if the sides of the prism were not definite boundaries but gradations in velocity, provided that velocities decrease at least to the value indicated somewhere within the prism. The velocity of 7.0 km/sec is therefore to be regarded as an upper limit to the minimum velocity. It seems likely that the 7.5 km/sec velocity determined by refraction below the axis of the Gregory Rift is a higher velocity upper surface of this zone, there being a gradual decrease in velocity with depth down to the low velocity suggested by our prism model.

If this prism is a zone of partially molten material, the corresponding shear velocity would be about 3.9 km/sec. As might be expected this minimum shear velocity is lower

than the mean shear velocity of 4.2 km/sec obtained for the top 80 km of the mantle from surface wave dispersion along the AAE–NAI path (model EA1, Fig. 1). Such is consistent with a normal velocity lid to the anomalous zone on the eastern flank as discussed earlier. It is also consistent with the concept of a velocity gradation from normal mantle to the ultra-low velocity material. Models of this type which have been fitted to the dispersion data generally suggest a depth to the anomaly of some 50 km beneath the eastern flanks of the rift.

The picture therefore emerging is of an ultra-low velocity zone similar to that suggested by Knopoff et al. (1970) to account for surface-wave dispersion along the East Pacific rise. The surface wave data by itself cannot define the detailed structure of such a model. The significance of the array data is that it provides the first evidence for ultra-low velocities beneath the rift system.

CONCLUSIONS

There is evidence that the crust and sub-Moho velocities do not depart significantly from those for the AFRIC model, except along the axis of the Gregory Rift. Here both surface wave dispersion and the apparent velocities of regional events indicate that the anomalous crust is of very limited lateral extent. The array data demonstrates that normal crust and sub-Moho material exist in proximity to the anomalous structure of the rift axis.

The apparent velocities from teleseismic events suggest that the boundary between normal and anomalous material extends into the mantle with a significant thinning of the anomalous mantle zone away from the rift. The preferred mantle model is of a zone of low velocity material which cuts into the crust along the axis of the Gregory Rift, but thins to run below normal mantle as we pass away from the Rift itself. It seems likely that the 7.5 km/sec velocity determined by refraction below the axis of the Gregory Rift is a higher velocity upper surface of this zone, there being a gradual decrease in velocity with depth down to some 7.0 km/sec or possibly lower. The lower surface of the anomaly may contribute to the thinning but no data is available to define its structure.

The Gregory Rift would therefore appear to lie along the junction of two blocks of thinning shield lithosphere separated by a narrow zone of anomalous material. The 6.4 km/sec velocity of the axial crust may represent some magmatic intrusion. There is a clear similarity between the crustal structure along the Rift axis and that of Iceland (Båth, 1960; Pálmason, 1970). The similarity of the mean compressional velocity of the mantle beneath East Africa to that beneath Iceland, and the suggestion of an ultra-low velocity zone similar to that suggested beneath the East Pacific rise, would seem to indicate a mantle structure similar to that beneath ocean ridges.

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HEAT FLOW MEASUREMENTS IN EAST AFRICA

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(Received February 1, 1972)

ABSTRACT

Von Herzen, R.P., 1972. Heat flow measurements in East Africa. In: R.W. Girdler (Editor), *East African Rifts. Tectonophysics*, 15(1/2): 179.

All heat-flow measurements in East Africa until 1971 have been obtained from relatively deep rift valley lakes by oceanographic techniques. Twenty values from Lake Malawi range over more than an order of magnitude (Von Herzen and Vacquier, 1967), although the average is similar to that obtained from South African shield measurements, about 1.1 μcal.cm^{-2}.sec^{-1} (H.F.U.). The values appear systematically distributed geographically over the lake area, with a possible correlation with localized tectonic activity. Except for one high value (3.5 H.F.U.), values from 12 locations in Lake Tanganyika range from low to normal (0.4–1.4 H.F.U.), with no apparent systematic geographical distribution (Degens et al., 1971). Similarly, some recent values from Lake Kivu have a range of 0.4–4.4 H.F.U., but most are above normal. The higher heat flow in Lake Kivu seems consistent with the recent volcanic activity in this region.

The effects of the local environment, particularly high sedimentation rates and past changes in lake water temperature, have probably affected many of the measurements. An unfavorable combination of these effects at some stations may imply equilibrium values which are 2 to 3 times larger than those measured. These could make some values comparable to high values measured on oceanic rifts. However, other geophysical measurements, particularly gravity, imply a continental rather than oceanic crustal structure beneath the lakes, suggesting that the tectonic mechanisms responsible for the anomalous heat flux are different from those thought to be associated with oceanic rifts, i.e., cooling of a spreading lithospheric slab.

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